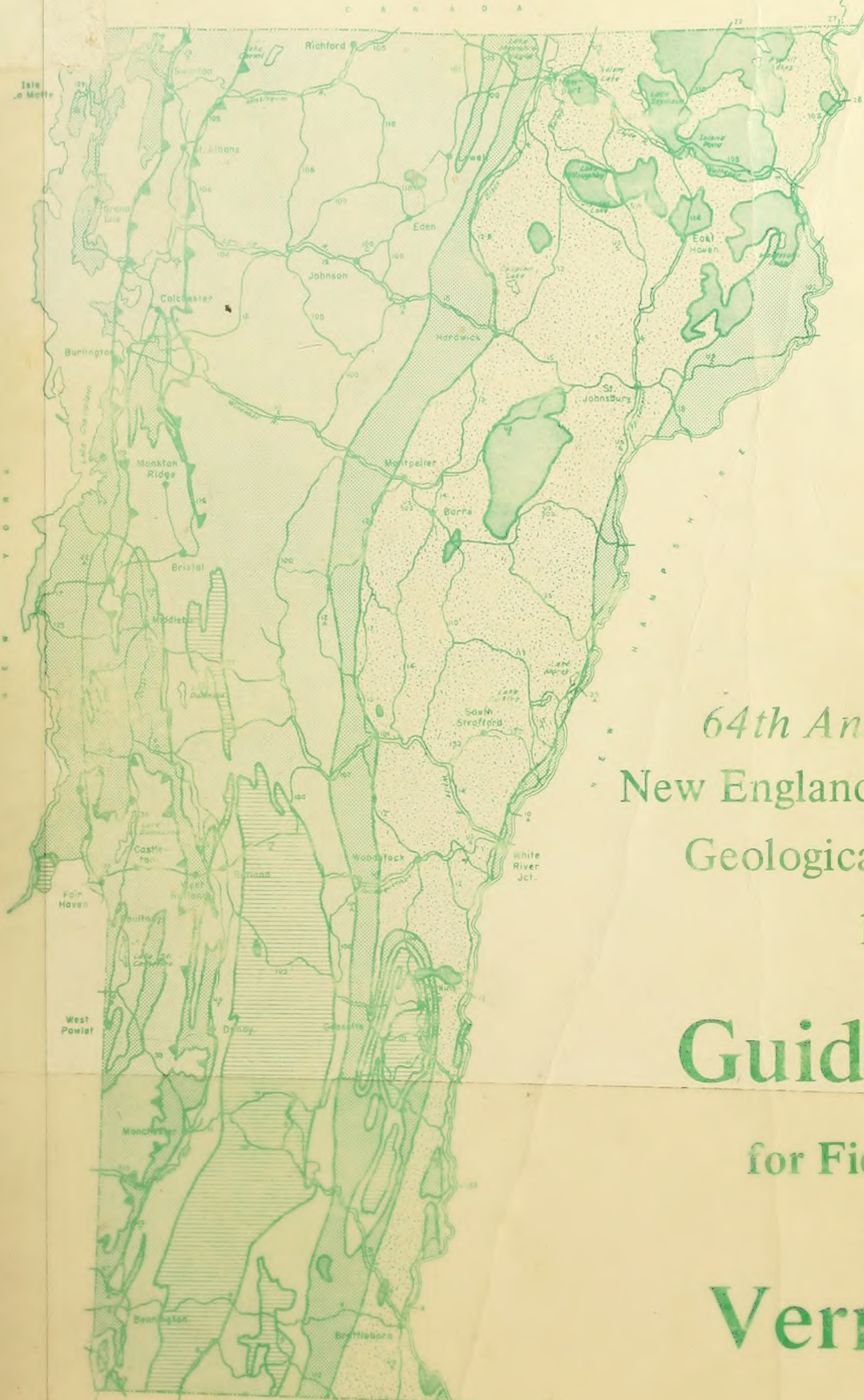


64th Annual Meeting

New England Intercollegiate Geological Conference

1972

Guidebook

for Field Trips

in

Vermont

New England Intercollegiate Geological Conference
64th Annual Meeting

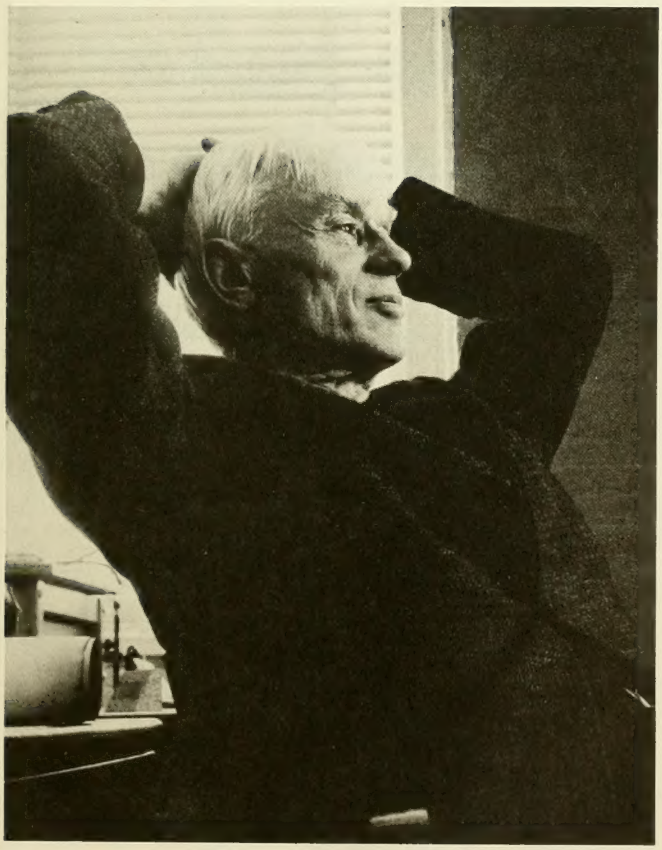
Guidebook for Field Trips in Vermont

October 13,14,15, 1972
Burlington, Vermont

Editors
Barry L. Doolan
Rolfe S. Stanley
University of Vermont

To Charles G. Doll

In deepest appreciation of his devotion and contributions to the understanding of Vermont Geology, this volume is affectionately dedicated by his many associates and friends.



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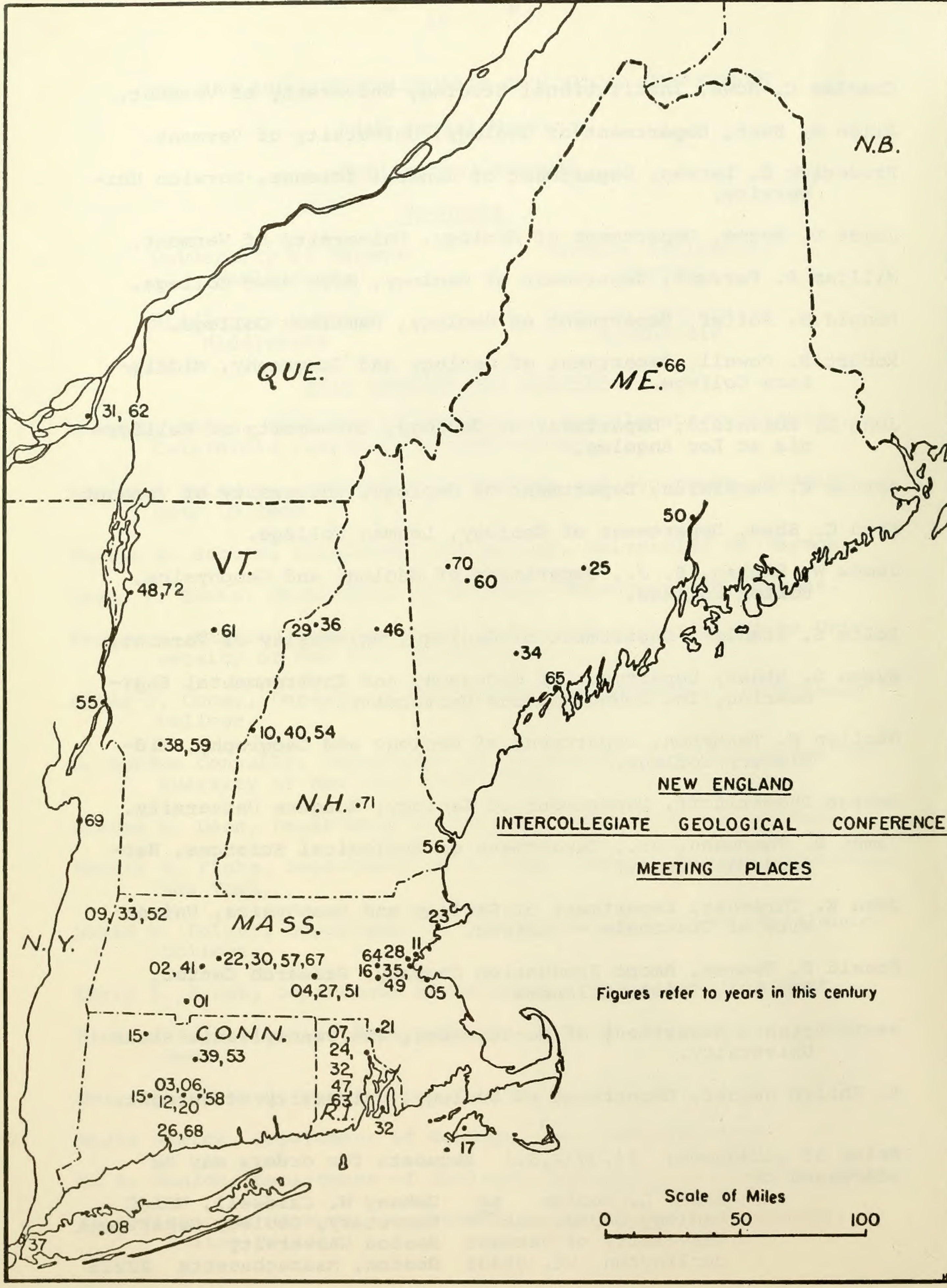
NEW ENGLAND
INTERCOLLEGIATE GEOLOGICAL CONFERENCE
MEETING PLACES
Figures refer to years in this century
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NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

CHRONOLOGICAL SUCCESSION OF MEETINGS

1.	1901	Westfield River Terrace, Mass.	Davis
2.	1902	Mount Tom, Mass.	Emerson
3.	1903	West Peak, Meriden, Conn.	Rice
4.	1904	Worcester, Mass.	Emerson
5.	1905	Boston Harbour and Nantasket	Johnson, Crosby
6.	1906	Meriden to East Berlin, Conn.	Gregory
7.	1907	Providence, R.I.	Brown
8.	1908	Long Island, N.Y.	Barrell
9.	1909	North Berkshires, Mass.	Cleland
10.	1910	Hanover, N.H.	Goldthwait
11.	1911	Nahant and Medford, Mass.	Lane, Johnson
12.	1912	Higby-Lamentation Blocks	Rice
13.	1915	Waterbury to Winsted, Conn.	Barrell
14.	1916	Blue Hills, Mass.	Crosby, Warren
15.	1917	Gay Head & Martha's Vineyard	Woodworth, Wigglesworth
16.	1920	Lamentation & Hanging Hills	Rice, Foye
17.	1921	Attleboro, Mass.	Woodworth
18.	1922	Amherst, Mass.	Antevs
19.	1923	Beverly, Mass.	Lane
20.	1924	Providence, R.I.	Brown
21.	1925	Waterville, Maine	Perkins
22.	1926	New Haven, Conn.	Longwell
23.	1927	Worcester, Mass.	Perry, Little, Gordon
24.	1928	Cambridge, Mass.	Billings, Bryan, Mather
25.	1929	Littleton, N.H.	Crosby
26.	1930	Amherst, Mass.	Loomis, Gordon
27.	1931	Montreal, Quebec	O'Neill, Graham, Clark, Gill, Osborne, McGerrigle
28.	1932	Providence-Newport, R.I.	Brown
29.	1933	Williamstown, Mass.	Cleland, Perry, Knopf
30.	1934	Lewiston, Maine	Fisher, Perkins
31.	1935	Boston, Mass.	Morris, Pearsall, Whitehead
32.	1936	Littleton, N.H.	Billings, Hadley, Cleaves, Williams
33.	1937	New York City & Duchess Co.	O'Connell, Kay, Fluhr, Hubbert, Balk
34.	1938	Rutland, Vt.	Bain
35.	1939	Hartford & Conn. Valley	Troxell, Flint, Longwell, Peoples, Wheeler
36.	1940	Hanover, N.H.	Goldthwait, Denny, Shaub, Hadley, Bannerman, Stoiber
37.	1941	Northampton, Mass.	Balk, Jahns, Lochman, Shaub, Willard
38.	1946	Mt. Washington, N.H.	Billings
39.	1947	Providence, R.I.	Quinn
40.	1948	Burlington, Vt.	Doll
41.	1949	Boston, Mass.	Nichols, Billings, Schrock, Currier, Stearns
42.	1950	Bangor, Maine	Trefethen, Raisz
43.	1951	Worcester, Mass.	Lougee, Little
44.	1952	Williamstown, Mass.	Perry, Foote, McFadyen, Ramsdell
45.	1953	Hartford, Conn.	Flint, Gates, Peoples, Cushman, Aitken, Rodgers, Troxell
46.	1954	Hanover, N.H.	Elston, Washburn, Lyons, McKinstry, Stoiber, McNair, Thompson
47.	1955	Ticonderoga, N.Y.	Rodgers, Walton, MacClintock, Bartolome
48.	1956	Portsmouth, N.H.	Novotny, Billings, Chapman, Bradley, Freedman, Stewart
49.	1957	Amherst, Mass.	Bain, Johannson, Rice, Stobbe, Woodland, Brophy, Kierstead, Webb, Shaub, Nelson
50.	1958	Middletown, Conn.	Rosenfeld, Eaton, Sanders, Porter, Lungren, Rodgers
51.	1959	Rutland, Vermont	Zen, Kay, Welby, Bain, Theokritoff, Osberg, Shumaker, Berry, Thompson
52.	1960	Rumford, Me.	Griscom, Milton, Wolfe, Calwell, Peacor
53.	1961	Montpelier, Vt.	Doll, Cady, White, Chidester, Matthews, Nichols, Baldwin, Stewart, Dennis
54.	1962	Montreal, Quebec	Gill, Clark, Kranck, Stevenson, Stearn, Elson, Eakins, Gold
55.	1963	Providence, R.I.	Quinn, Mutch, Schafer, Agron, Chapple, Feininger, Hall
56.	1964	Chestnut Hill, Mass.	Skehan
57.	1965	Brunswick, Me.	Hussey
58.	1966	Katahdin, Me.	Caldwell
59.	1967	Amherst, Mass.	Robinson, Drake
60.	1968	New Haven, Conn.	Orville
61.	1969	Albany, N.Y.	Bird
62.	1970	Rangeley Lakes, Me.	Boone
63.	1971	Concord, N.H.	Lyons, Stewart
64.	1972	Burlington, Vt.	Doolan, Stanley

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FORWARD

> "Largely through....(the publication of) the new geologic map of the state, a widespread active interest has been created among geologists who will come to Vermont to study and make comparisons with the geology of already classic areas elsewhere." (Doll, 1962, p.11)

These words written by Charles G. Doll in the 1960-62 Biennium Report of the State Geologist were published a year after the last meeting of the N.E.I.G.C. in Vermont, a meeting hosted by the Vermont Geological Survey in celebration of the publication of the Centennial Geologic Map of Vermont.

Charles Doll's perceptive insights of the impact of the new map on the geologic community have proven to be modestly correct. In the eleven years since the publication of the Centennial Map, Vermont geology has been undergoing active reexamination to answer the seemingly never-ending problems of the complex geologic history of the northern Appalachians and, more recently, to relate Vermont to the history of continental drift and plate tectonics in the North Atlantic.

Also, in the last eleven years, through the effort of Charles Doll, the Vermont Geological Survey has initiated and completed a surficial mapping program culminating in the publication of the Surficial Geologic Map of Vermont in 1970, and initiated an Environmental Geology mapping program actively in progress for the past two years.

With these accomplishments and activities clearly in mind, the editors, the many guidebook contributors, and numerous workers have striven to compile a guidebook which reflects the diversity of subject and areal extent of Vermont's complex geology.

In meeting these goals and in anticipation of active participation of large numbers of professionals, students, and teachers at this conference, a record number of trips have been organized (24) in Bedrock Geology, Environmental Geology, Glacial Geology, Lake Studies, and Paleontology. These trips have been alphabetically arranged under the appropriate headings in the following pages for ease of reference and continuity of subject matter.

To accommodate more active participation by all those attending the conference, the editors sought (and gratefully received) cooperation from the contributors to publish the guidebook ahead of schedule so that it can be in the hands of participants before they attend the conference. Since many of these participants will be students and professionals unfamiliar with many aspects of Vermont geology, we have compiled in the immediately following introductory pages, pertinent maps and tables from Vermont Geological Survey publications for their perusal. A complete listing of Vermont Geological Survey publications is appended at the rear of this guidebook for those wishing to obtain additional, more complete information on various aspects of Vermont geology before the conference.

By including all this material under one cover to supplement the very fine array of papers by the contributors we sincerely hope that this guidebook will also be of use to high schools, laymen, and many university-organized field trips in the years to come.

Acknowledgements

With pleasure we gratefully acknowledge the significant contributions, generous suggestions and efforts of the field trip leaders and authors. The early publication of a guidebook with thirty-six contributors, nearly 500 pages, and twenty-four field trips is indeed a tribute to their cooperation. A great deal of this credit however must also go to two extremely dedicated individuals who prepared the papers for publication, handled the numerous logistical problems, and offered many suggestions. Margaret Newton, department secretary, proofread the entire guidebook and typed most of its pages. Terry Frank, our "NEIGC secretary" completely revamped the NEIGC mailing list, organized the trip lists and handled all the correspondence and finances of this conference. To both of them thanks for a job well done ! Miscellaneous drafting chores and preparation of many of the guidebook figures by Sally Rising and Thelma Barton, University of Vermont geology students, are also gratefully acknowledged.

Special thanks go to Art Huse of the UVM Geology Department for printing and layout of the "chapter headings" of this guidebook.

Finally we acknowledge John Wiley and Sons for permission to reproduce Figures 14-1, 14-5 and Table 14-1 from Zen, E. and others, Studies of Appalachian Geology: Northern and Maritime, for John Rosenfeld's trip (B-7), and the Vermont Geological Survey for permission to print figures, maps, and tables from earlier publications.

This guidebook has been printed and bound by the University of Vermont Print Shop.

Barry L. Doolan
Rolfe S. Stanley
Department of Geology
University of Vermont
Editors


FIELD TRIP LOCALITIES
1972 N.E.I.G.C. 64th ANNUAL MEETING
BURLINGTON, VERMONT
LEGEND
Bedrock
Environmental
Glacial
Lake Studies
Paleontology
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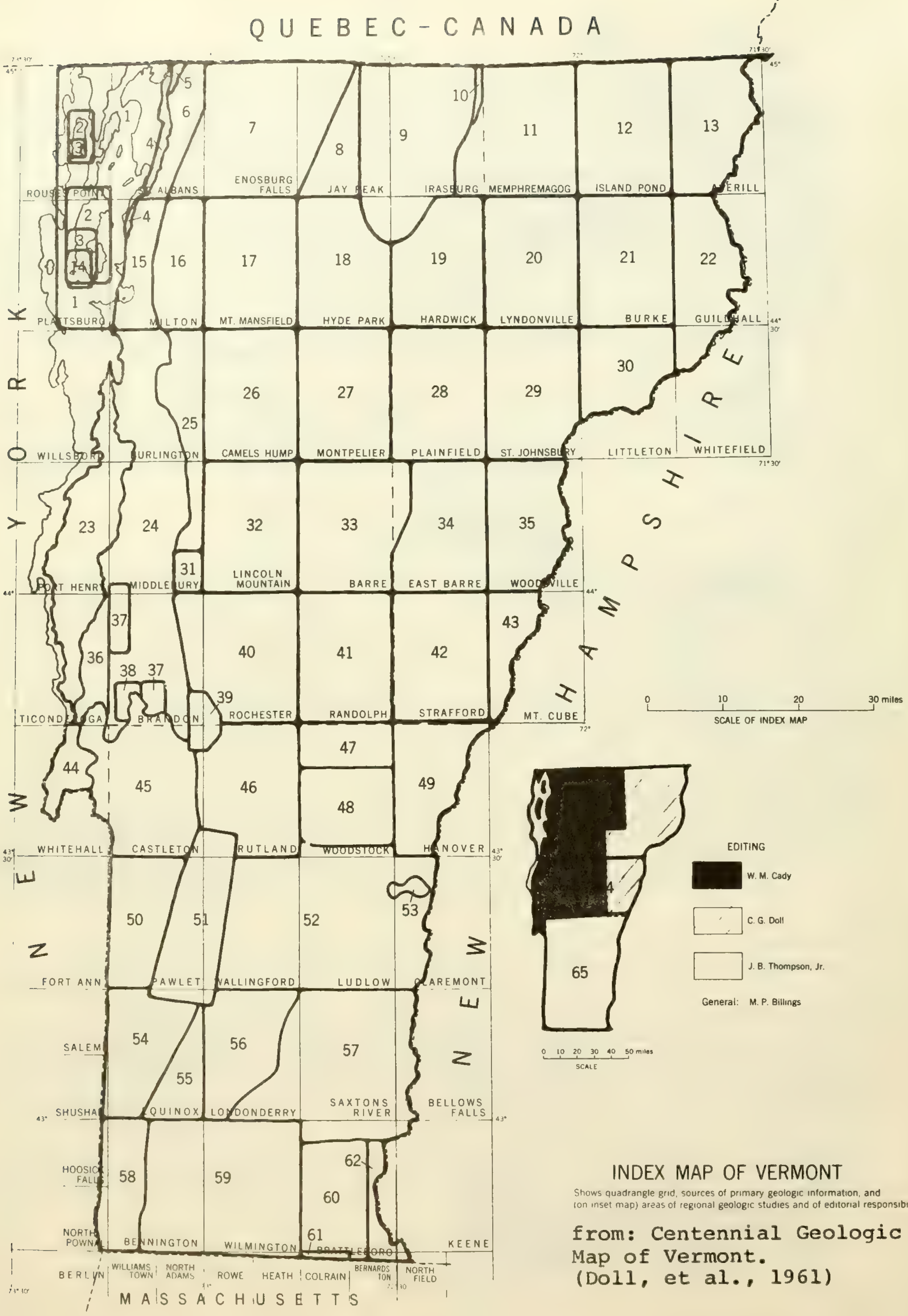
Essex Junction
Burlington
St Johnsbury
Montpelier
Barre
Middlebury
Rutland
Springfield
Bennington
Brattleboro



INDEX MAP OF VERMONT

Shows quadrangle grid, sources of primary geologic information, and (on inset map) areas of regional geologic studies and of editorial responsibility

from: Centennial Geologic Map of Vermont. (Doll, et al., 1961)

PRIMARY GEOLOGIC INFORMATION

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62 Moore, G E, Jr, 1949, Structure and metamorphism of the Keene Brattleboro area, New Hampshire-Vermont Geol Soc American Bull, v 60, p 1613-1669, **also**, Balk, Robert, 1956, Bedrock geology of the Bernardston quadrangle, Massachusetts-Vermont U S Geol Survey Geol Quad Map GQ 90 and Bedrock geology of the Massachusetts portion of the Northfield quadrangle, Massachusetts, New Hampshire, and Vermont U S Geol Survey Geol Quad Map GQ-92

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64 White, W S, and Johns, R H, 1950, Structure of central and east-central Vermont Jour Geology, v 58, p 179-220

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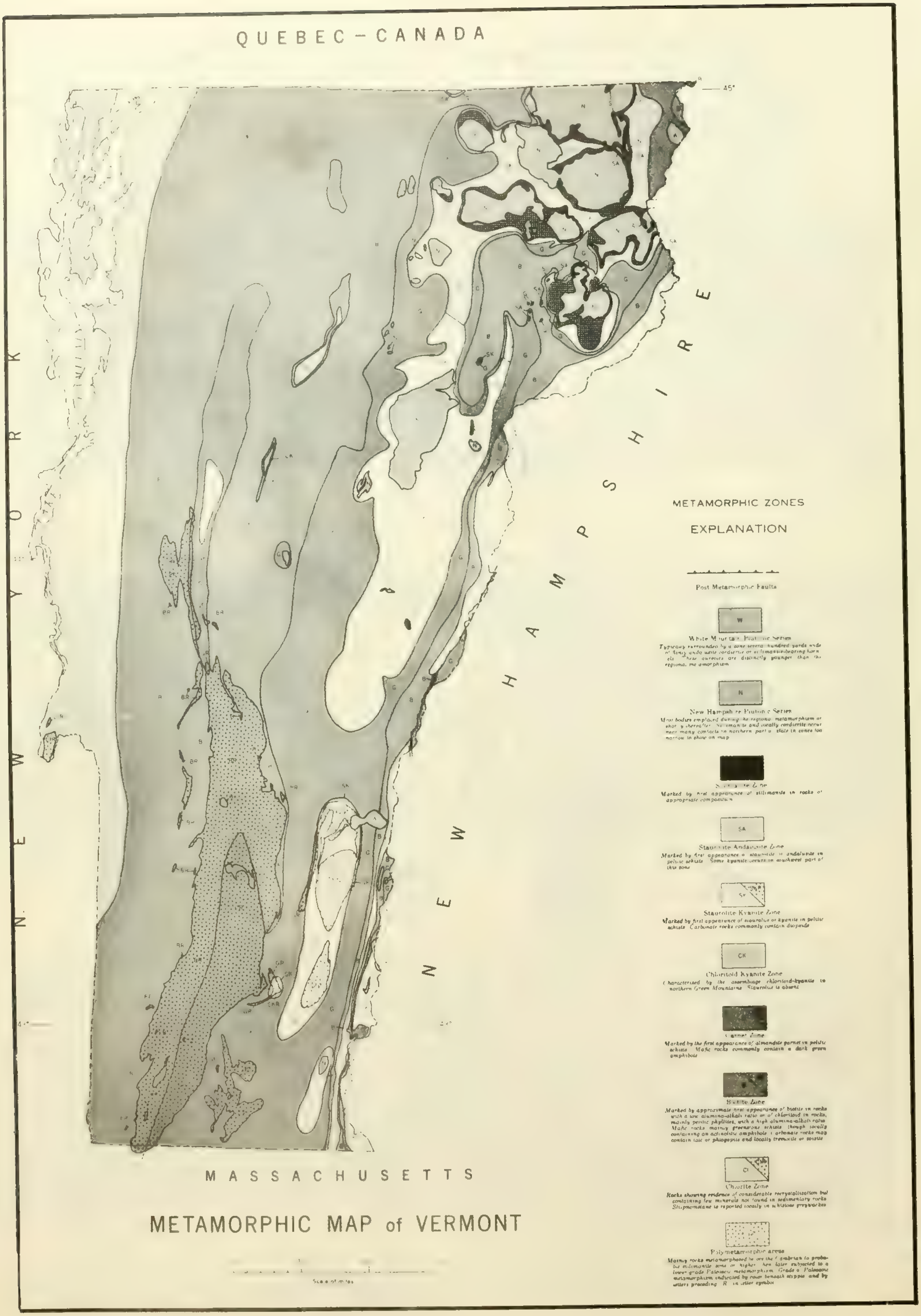
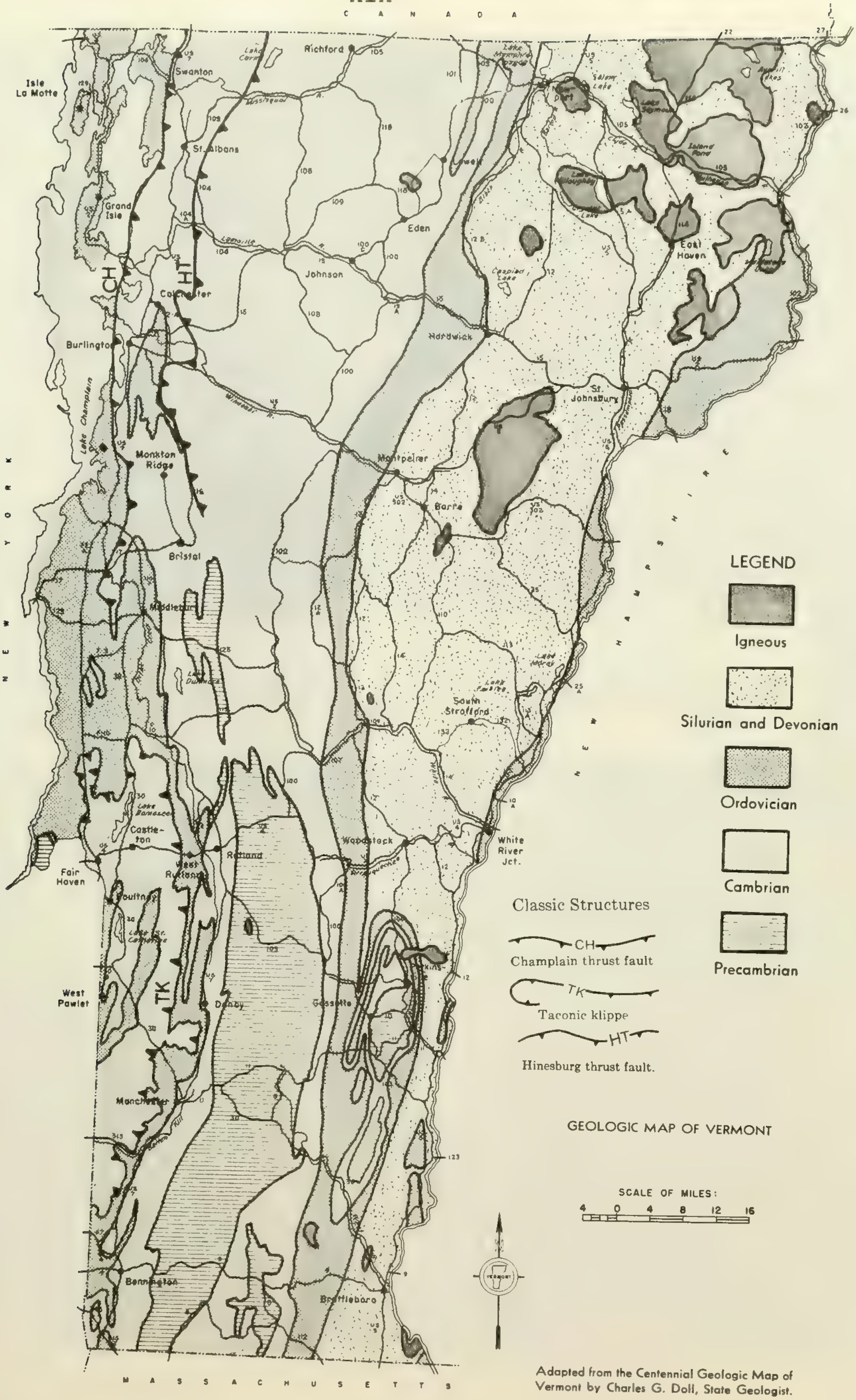
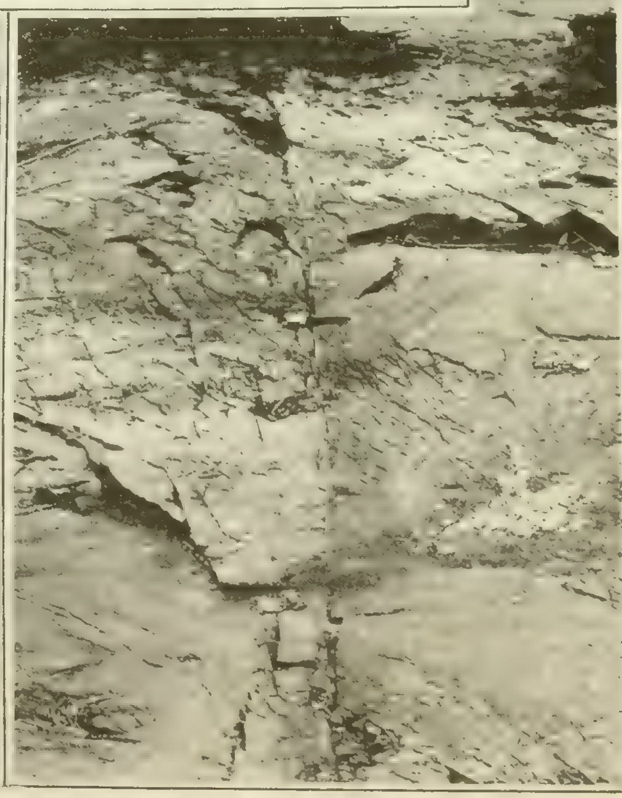


METAMORPHIC MAP of VERMONT


CANADA
NEW YORK
NEW HAMPSHIRE
MASSACHUSETTS
Isle La Motte
Swanton
Richford
St. Albans
Grand Isle
Eden
Johnson
Colchester
Burlington
Lake Champlain
Hardwick
East Haven
St. Johnsbury
Montpelier
Barre
Monkton Ridge
Bristol
Middlebury
South Strafford
Castleton
Fair Haven
West Rutland
Rutland
Woodstock
White River Jct.
Poultney
West Pawlet
Danby
Manchester
Bennington
Brattleboro
CH
HT
TK
LEGEND
Igneous
Silurian and Devonian
Ordovician
Cambrian
Precambrian
Classic Structures
Champlain thrust fault
Taconic klippe
Hinesburg thrust fault.
GEOLOGIC MAP OF VERMONT
SCALE OF MILES:
4 0 4 8 12 16
Adapted from the Centennial Geologic Map of Vermont by Charles G. Doll, State Geologist.

bedrock geology



Bedrock Geology Cover page: Upper: Champlain overthrust at Rock Point, Lake Champlain, Vermont; see Stanley and Sarkisian (Trip B-4, this guidebook). Photo by Terry Frank. Lower left: Detail of flow structure at the hinge of Scotch Hill syncline within dolostone bed (actual size approximately 1.5 x 2 inches). See Locality 2, Trip B-3; Barry Voight (this guidebook). Lower right: Shelburne fishing access area, Shelburne, Vermont; detail of sinistral fault and associated fractures; see Stanley and Sarkisian (Trip B-4, this guidebook).

Trip B-1

STRATIGRAPHY OF THE EAST FLANK OF THE GREEN MOUNTAIN ANTICLINORIUM, SOUTHERN VERMONT

by

James W. Skehan, S.J.* and J. Christopher Hepburn*

INTRODUCTION

The Green Mountain anticlinorium in southern Vermont has an exposed core of Precambrian gneisses overlain to the east and west by metamorphosed Paleozoic rocks. The rocks of the west limb of the anticlinorium are chiefly quartzites and carbonates of a miogeosynclinal sequence. The east limb of the anticlinorium consists of a eugeosynclinal sequence of schists and gneisses from (?)Cambrian through Lower Devonian age. The purpose of the present field trip is to examine the stratigraphy of these schists and gneisses. A roughly west-to-east section across portions of the Wilmington and Brattleboro quadrangles (Fig. 1) will be followed.

The earliest geological mapping in the area was done by E. Hitchcock and others during the compilation of the Geology of Vermont (E. Hitchcock *et al.*, 1861). Hubbard (1924), Prindle and Knopf (1932), Richardson (1933), and Richardson and Maynard (1939) studied portions of the area. Thompson (1950) and Rosenfeld (1954), working in the Ludlow and Saxtons River quadrangles respectively, started the comprehensive detailed mapping of southern Vermont. Detailed geological mapping of the field trip area has been compiled by the authors (Fig. 1). The Wilmington-Woodford area was mapped by Skehan (1953, 1961), and the Brattleboro area by Hepburn (1972b).

Most of the stratigraphic units of the east limb of the Green Mountain anticlinorium in southern Vermont can be traced directly from the Wilmington-Brattleboro area to their type localities further north in Vermont or to the south in Massachusetts. A few can be traced into fossiliferous strata. Currently a number of workers (see for example Hatch, Osberg, and Norton, 1967; Hatch, 1967; Hatch, Schnabel and Norton, 1968) are tracing many of these units and their correlatives southward through western Massachusetts and western Connecticut.

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STRATIGRAPHY

The stratigraphy of the field trip area is briefly summarized below. See Skehan (1961) and Hepburn (1972b) for more complete descriptions.

Wilmington Gneiss

The Wilmington Gneiss named by Skehan (1961) is of uncertain stratigraphic position. It may be Precambrian in age, resembling as it does the microcline gneiss sequence of the Mt. Holly Complex of the Green Mountain core. On the other hand the apparently conformable relationship immediately beneath the Hoosac and Tyson Formations along their eastern contact (Fig. 1) suggests strongly the possibility that the Wilmington Gneiss may be of Cambrian age. The complex and unexplained relationships of the Wilmington Gneiss to the members of the Cavendish Formation of Doll *et al*. (1961) along the western contact make a decision as to the age of the Wilmington Gneiss impossible at this time.

The Wilmington Gneiss consists of a medium to very coarse-grained, well-banded, somewhat foliated biotite-epidote-quartz-microcline augen gneiss. The microcline is gray to pink and occurs as lenticular augen and flaser in which the average long diameter is about 7mm. Locally the augen may reach 8 in. in length and are usually flattened into the plane of the foliation. Quartz rods and linearly aligned streaks of biotite are a common feature of the Wilmington Gneiss.

The Wilmington Gneiss may be the correlative of the Bull Hill Gneiss of Doll *et al*. (1961), an exposure of which is only one mile north of and on line with the northernmost exposure of the Wilmington Gneiss of the Wilmington quadrangle (Skehan, 1961, Pl. I).

Tyson Formation

The Tyson Formation, named by Thompson (1950), is recognized in this area only as discontinuous lenses of fine to coarse-grained, schistose, white to blue quartz-pebble conglomerate; fine to coarse-grained gray, buff and pink microcline-pebble and coarse-grained albite-pebble conglomerate and conglomeratic schist; and thin-bedded quartzite and dark biotite-muscovite-quartz schist.

Hoosac Formation

The Hoosac Formation (Hoosac Schist of Pumpelly et al., 1894) consists of gray, brown and black, medium to coarse-grained muscovite-biotite-albite-quartz schists locally containing variable amounts of chlorite, muscovite, paragonite and garnet. Rocks containing appreciable garnet commonly weather to a mottled rusty color. Albite megacrysts 2-15 mm. in diameter are characteristic of the formation, which is distinguished from the overlying Pinney Hollow Formation by the presence of more abundant albite megacrysts, its color, and its generally coarser and more granular texture.

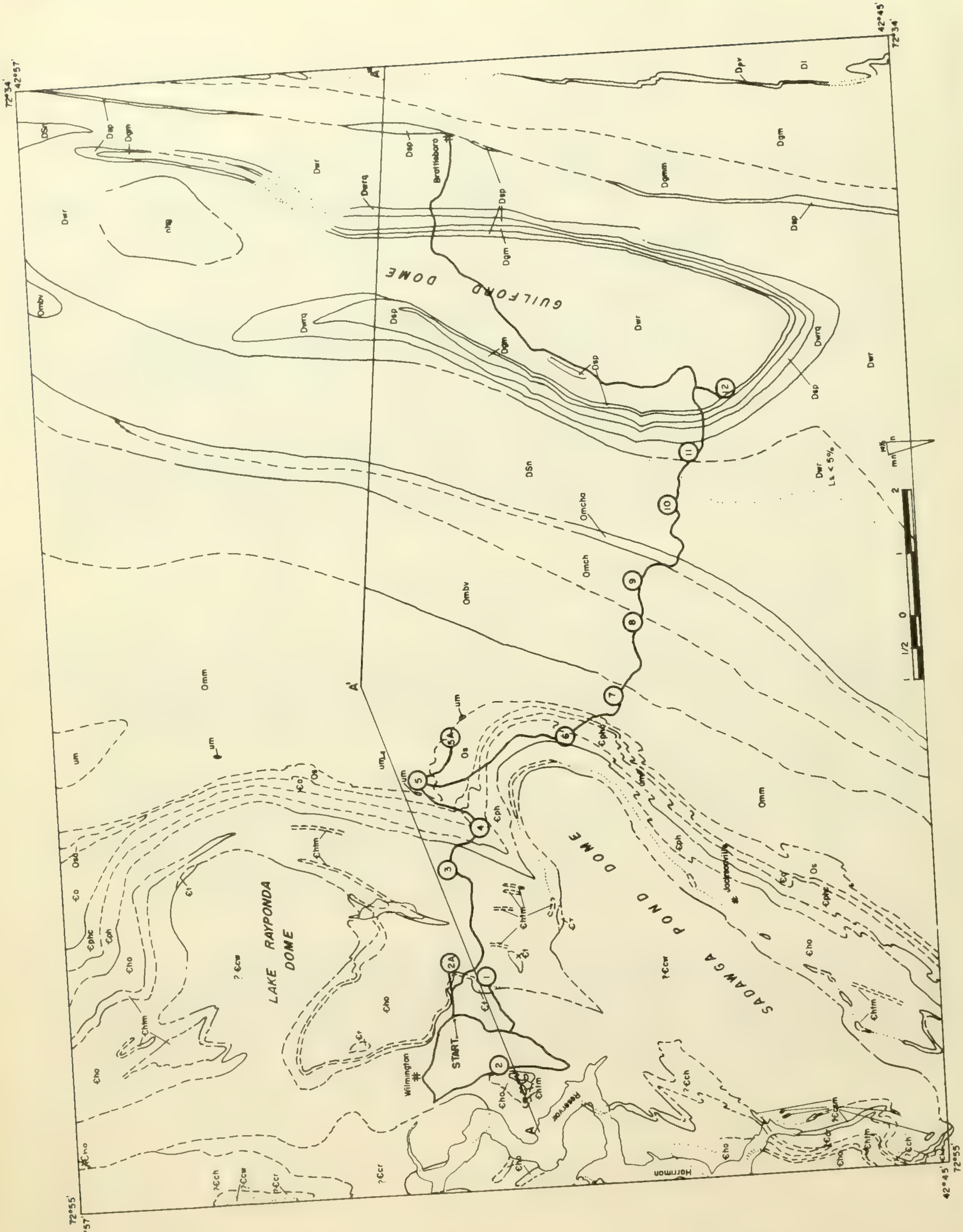
The Turkey Mountain Member of the Hoosac Formation (named by Rosenfeld, 1954) is typically a dense dark green to black amphibolite commonly characterized by rounded to sub-angular white, gray, green or dark brown "amygdules" composed of quartz and albite commonly with included epidote, hornblende and garnet.

Pinney Hollow Formation

The Pinney Hollow Formation named by Perry (1928) is characteristically a pale to dark green well-foliated chlorite-muscovite-(paragonite)-chloritoid-garnet-quartz schist. Epidote-albite-hornblende amphibolite including amygdaloidal amphibolite is interbedded with the chlorite schist.

The Chester Amphibolite named by Emerson (1898b) and mapped as a separate formation by Skehan (1961), is here considered as a member of the Pinney Hollow Formation, following the usage of Doll et al. (1961). The Chester Amphibolite is mapped as the first thick sequence of amphibolites above the dominantly green Pinney Hollow chlorite schists and immediately below the black schists of the Ottauquechee Formation.

The Chester Amphibolite is characteristically a banded, well-foliated epidote-chlorite-albite-hornblende schist containing thin beds of dark gray to black muscovite-quartz schist and green chlorite-muscovite-garnet-quartz schist.


LAKE RAYPONDA DOME
SADAWGA POND DOME
GUILFORD DOME
START
Wilmington
Brattleboro
Jacksonville
Harriman Reservoir

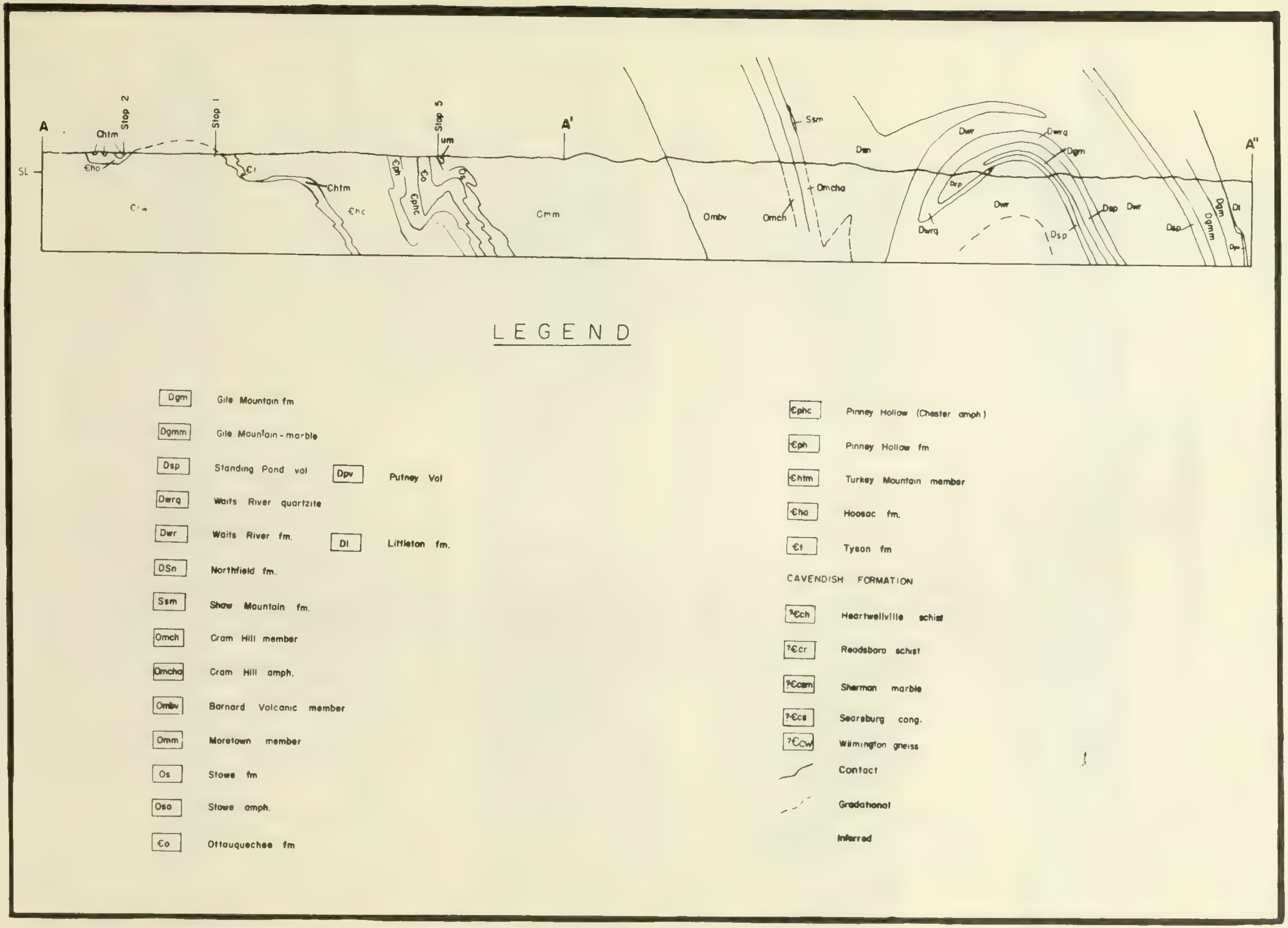


FIGURE I

GEOLOGIC MAP OF THE WILMINGTON - BRATTLEBORO AREA

Ottauquechee Formation

The Ottauquechee Formation named by Perry (1928) is characteristically a rusty-weathering uniform sequence of black sulfide-bearing, muscovite-garnet-(chlorite)-quartz schist; sulfide-bearing feldspathic quartzite and vitreous black quartzite; and feldspathic biotite-quartz schist.

Stowe Formation

The Stowe Formation named by Cady (1956) is identical to the Pinney Hollow Formation proper. In the northern part of the Wilmington quadrangle a thick amphibolite unit similar to the Chester Amphibolite forms the base of the sequence and has been mapped separately (Osa).

Missisquoi Formation

Following Doll et al. (1961) the Missisquoi Formation consists of the rocks lying between the Stowe Formation and the Ordovician-Silurian unconformity at the base of the Shaw Mountain Formation. It is continuous from the Wilmington-Brattleboro area to northern Vermont where it was defined as the Missisquoi Group by Richardson and Cabeen (1923). The Missisquoi Formation is subdivided into three members.

Moretown Member

The Moretown Member, defined by Cady (1956) as the Moretown Formation, is characterized by distinctive interlayered quartz-mica schists, mica schists, and quartz-rich granulites. Typically, light gray granulites and quartz-rich schists are interlaminated with light to dark gray micaceous laminae on a scale of 2 mm. to a few centimeters in thickness. Where the lamination is continuous and abundant on a fine scale, the rock has a distinctive "pinstriped" appearance. At least some of the thin interlayering of the micaceous bands is of secondary origin. Garnet and chlorite porphyroblasts are common and pyrite cubes as large as 1.5 cm. are not an uncommon accessory. Fasciculitic chlorite and biotite sprays after an amphibole are seen on some schistosity surfaces. Interlayers of amphibolite and garnet amphibolite account for up to 20 percent of the member.

SYSTEM		Wilmington-Brattleboro Area		Western Mass. (Hatch *et al.*, 1968 and 1970)	Western Mass. (Emerson, 1898a, 1898b, and 1917)	New Hampshire (Billings, 1956; Thompson *et al.*, 1968)
DEVONIAN		Gile Mountain Fm.		Waits River Formation	Leyden Argillite	Littleton Formation; Erving Member
DEVONIAN		Putney Volcanics		Waits River Formation	Conway Amphib.	Littleton Formation
DEVONIAN	LOWER	? — ? — ? — ?				
DEVONIAN	LOWER	Standing Pond Vol.		Waits River Formation / Goshen Formation	Conway Schist	Littleton Formation
DEVONIAN	LOWER	Waits River Fm.		Waits River Formation / Goshen Formation	Conway Schist	Littleton Formation
SILURIAN		Northfield Formation		Goshen Formation	Goshen Schist	Fitch Formation
SILURIAN		Shaw Mt. Fm.		Russell Mt. Fm.	Goshen Schist	Clough Quartzite
ORDOVICIAN	M	Missisquoi Fm.	Cram Hill Mbr.	Hawley Formation	Goshen Schist	Partridge Fm.
ORDOVICIAN		Missisquoi Fm.	Barnard Volcanics Member	Hawley Formation	Hawley Schist	Ammonoosuc Vol.
ORDOVICIAN	L	Missisquoi Fm.	Moretown Member	Moretown Formation	Savoy Schist	Albee Formation
CAMBRIAN		Stowe Formation		Rowe Schist	Savoy Schist	
CAMBRIAN		Ottauquechee Fm.		Rowe Schist	Savoy Schist	
CAMBRIAN			Chester Amphib.	Rowe Schist	Chester Amphib.	
CAMBRIAN		Pinney Hollow Fm.		Rowe Schist	Rowe Schist	
CAMBRIAN		Hoosac Formation		Hoosac Fm.	Hoosac Schist	
?CAMBRIAN		Cavendish	Heartwellville			
?CAMBRIAN		Cavendish	Readsboro			
?CAMBRIAN		Cavendish	Sherman Marble			
?CAMBRIAN		Cavendish	Searsburg Cgl.			
?CAMBRIAN		Cavendish	Wilmington Gn.			
PRECAMBRIAN		Mt. Holly Complex Stamford Granite Gneiss		Stamford Granite Gneiss	Becket Gneiss	

Figure 2. Correlation Chart of the Wilmington-Brattleboro Area, Vermont.

Barnard Volcanic Member

The Barnard Volcanic Member includes a wide variety of rocks, but three general types are most abundant: non-porphyritic amphibolite; porphyritic amphibolite with numerous feldspar megacrysts; and light-colored, felsic schist and gneiss. The felsic rocks make up 35 to 50 percent of the member.

Cram Hill Member

The Cram Hill Member, first designated as the Cram Hill Formation by Currier and Jahns (1941), is correlative with fossiliferous late Middle Ordovician rocks at Magog, Quebec (Berry, 1962). The member consists largely of fine-grained black phyllite and schist that weathers a rusty-brown from finely disseminated pyrite or pyrrhotite. Thin, fine-grained, black quartzite beds are not uncommon. Biotite and pyrite are the common porphyroblasts, and garnet may be present. Thin amphibolite layers are common. A zone at the top of the member in which amphibolite predominates has been separated on Figure 1 as Omcha.

Shaw Mountain Formation

In the Brattleboro quadrangle the Silurian Shaw Mountain Formation (Currier and Jahns, 1941) occurs in only three thin lenses to the east of the regional unconformity at the top of the Missisquoi Formation. White to light brown weathering quartzite and quartz-pebble conglomerate occur at the base of the formation. These rocks grade upward into quartz-mica schist, mica schist, a coarse-grained hornblende fasciculite schist, and amphibolite.

Northfield Formation

The Northfield Formation, named the Northfield Slate by Currier and Jahns (1941), is a uniform sequence of gray to dark gray, graphitic quartz-muscovite schists with very conspicuous almandine porphyroblasts 1 to 2 mm. in diameter. Biotite is a common additional porphyroblast, as is staurolite at the appropriate grade. Crinkle lineations are commonly well developed in the schists. Thin quartzite beds are occasionally present but are not as abundant as they are along strike in the Goshen Formation of western Massachusetts (Hatch et al., 1968). A few punky-brown weathering impure marble beds similar to those in the overlying Waits River Formation are present.

Waits River Formation

The Waits River Formation (Currier and Jahns, 1941) consists of interbeds of three broad categories of rocks: impure marbles, various schists, and impure quartzites. The most distinctive rocks are the impure marbles, which weather to a punky-brown and have a friable surficial rind of non-calcareous minerals, largely quartz, left behind where the carbonate has weathered out. The impure marbles occur in beds a few inches to tens of feet thick. The percentage of these beds varies throughout the formation.

The schistose rocks are quite variable but generally weather dark gray to brown and contain quartz and muscovite with differing amounts of biotite, garnet, or carbonate. Zoisite may be an additional porphyroblast.

Thin beds of light gray feldspathic and micaceous quartzite occur throughout the formation. A narrow band adjacent to the Standing Pond Volcanics has been separated (Dwrq, Fig. 1) in which the impure quartzites are present to the exclusion of the other rocks.

Standing Pond Volcanics

The Standing Pond Volcanics (Doll, 1944, Standing Pond Amphibolite) consist mostly of black, massive to moderately foliated amphibolite and epidote amphibolite. Very coarse-grained garnet amphibolite is commonly present near the contact with the Waits River Formation. Minor amounts of brown-weathering schist, feldspathic quartzite and coticule are also present. The eastern band of the Standing Pond Volcanics (Fig. 1) is composed of plagioclase-biotite-quartz and plagioclase-biotite-hornblende-quartz granulite.

Gile Mountain Formation

In the area to be visited by this field trip the Gile Mountain Formation (Doll, 1944, Gile Mountain Schists) is metamorphosed to the kyanite-staurolite zone. At this grade of metamorphism the principal rocks in the Gile Mountain Formation are micaceous and feldspathic quartzites and mica schists. The quartzites weather light gray and contain variable amounts of quartz, feldspar, muscovite, and biotite. Garnet, hornblende, and ankerite are present in minor amounts. Kyanite, staurolite, and garnet porphyroblasts are common in the mica schist interbeds.

STRUCTURE

The axial trace of the regional Green Mountain anticlinorium (Skehan, this volume, Sunday Trip, Figs. 1 and 3) lies just to the west of the field trip area. Most of the units seen on the field trip dip moderately to steeply east and are part of the homoclinal sequence of the east limb of the anticlinorium (Fig. 1). The Lower Cambrian miogeosynclinal facies, well-developed on the western limb of the anticlinorium, can be traced to its southeast limb, where it appears to be cut off by the Hoosac thrust fault. The (?)Cambrian Cavendish Formation in the southern part of the area overlies these Lower Cambrian rocks and the Precambrian core as a result of westward thrusting. In the northern part of the area, the Cavendish may be separated from the Precambrian by an angular unconformity or by the continuation of the Hoosac thrust. Overlying the Cavendish Formation is the Cambrian and Ordovician sequence of metasediments and metavolcanics. The Lake Rayponda and Sadawga Pond domes (Fig. 1) have locally disrupted the eastern limb of the anticlinorium.

The Guilford dome lying east of the Green Mountain anticlinorium is part of a belt of domes that stretches from central Vermont to Connecticut, west of the Connecticut River. The Chester and Athens domes, just north of the field trip area, have exposed Precambrian rocks in their cores. The Siluro-Devonian Waits River Formation is exposed in the center of the Guilford dome. Large recumbent folds are present in the strata mantling these domes in eastern Vermont. The doubly-closed loop of the Standing Pond Volcanics (Fig. 1) outlines such a fold, the Prospect Hill recumbent fold. This recumbent fold had formed prior to the doming and had a NE-SW trending axis. The subsequent doming arched the axial surface of the recumbent fold, causing the hinges to plunge moderately NE and SW away from the roughly N-S axial trace of the dome.

A sequence of at least four minor fold stages (Hepburn, 1972a) has been worked out for the eastern portion of the field trip area. The following sequence of minor folds is inferred:

(1) Small isoclinal folds with a schistosity developed parallel to their axial surfaces.

(2) Tight to isoclinal folds related to the Prospect Hill recumbent fold. A schistosity is developed parallel to the axial surfaces of these folds in some of the metapelites.

(3) Open folds with a slip-cleavage developed parallel to their axial surfaces. This cleavage generally strikes northeast and dips moderately to steeply northwest. These folds may have formed during the development of the Guilford dome.

(4) One or more generations of open folds, warps, or buckles in the foliation that fold the slip-cleavage.

Metamorphism

The area was regionally metamorphosed during the Acadian Orogeny. At this time the Precambrian rocks in the exposed core of the Green Mountain anticlinorium, previously metamorphosed to a high degree during the Grenville Orogeny, were extensively retrograded. The rest of the field trip area was metamorphosed to the garnet zone, except for the Guilford dome where the staurolite-kyanite zone was reached, as seen at the last two stops.

Road Log for Trip
Friday, October 13, 1972

James W. Skehan, S.J. and J. Christopher Hepburn, Leaders

Primary references for this trip are:

Doll, et al., (1961) Centennial Geologic Map of Vermont, October, 1961 ($4.00).

Hepburn, J.C. (1972b) Geology of the Metamorphosed Paleozoic Rocks in the Brattleboro Area, Vermont, unpub. Ph.D. Thesis, Harvard University, 377 p.

Skehan, J.W., S.J. (1961) Geologic Map of the Wilmington-Woodford Area, from Bulletin 17, Vermont Geological Survey (25¢)

Skehan, J.W., S.J. (1961) The Green Mountain Anticlinorium in the Vicinity of Wilmington and Woodford, Vermont, Bulletin 17, Vermont Geological Survey, 159 p. ($3.00)

All references, except Hepburn (1972b) may be obtained from the State of Vermont, Department of Libraries, Montpelier, Vermont. Enclose payment with order.

Mileage

0.00 Assemble at the junction of Routes 9 and 100, 1.15 miles east of the center of Wilmington, Vermont in the parking lot of Coombs' Beaver Brook Sugar House (Fig. 1). Park south of the store out of traffic. West of Route 100 at this locality, large outcrops of Wilmington Gneiss can be seen and may be examined by those arriving early.

Departure time at 8:45 a.m. SHARP. Proceed to Stop 1 by going south on Route 100.

0.85 Turn left and proceed up the hill to Hubbard Hill Farm.

1.35 Stop 1. WILMINGTON GNEISS, TYSON AND HOOSAC FORMATIONS.

Park off the unpaved road near the farm buildings of Hubbard Hill Farm. To the northwest one may see the Wilmington Valley and the Haystack-Mount Snow Ridge. Proceed on foot uphill to the northeast toward the crest of the hill. The outcrops west of the hillcrest are those of the micaceous microcline augen Wilmington Gneiss. At this locality, between the well-developed Wilmington Gneiss and the typical micaceous albite schists of the

Mileage (cont'd)

Hoosac Formation, is a blue-quartz-bearing gneiss and schistose gneiss assigned to the Tyson Conglomerate (Skehan, 1961, pp. 65-66).

The structure is dominated by cascade folds in which the movement sense is such that the upper beds have moved easterly relative to the lower beds. Some outcrops on the eastern flank of this hill show well-developed cascade folds with amplitudes of 1 1/2 feet. Although a number of individual outcrops show shallow to steep westerly dips, the beds have an average dip to the east. The axial planes of the cascade folds dip at an average of 50° NW. Return to cars, turn around, and return 0.45 mile to Route 100.

1.75 Proceed south on Route 100.

1.77 Outcrops of rusty weathering, dark albite schist of the Hoosac Formation underlain by banded plagioclase gneiss of the Wilmington Gneiss on the east side of the highway.

1.95 Turn right off Route 100 south at the Dix School. Proceed southwest on Boyd Hill Road. This road crosses terrain underlain by the micaceous microcline augen gneisses of the Wilmington Gneiss Formation.

3.35 Stop 2. HOOSAC FORMATION, TURKEY MOUNTAIN MEMBER AND WILMINGTON GNEISS.

Park off the road wherever you can near two houses on the left and the barn on the right. Proceed to the hill by a path between the two houses. Climb up the hill examining the Wilmington Gneiss in the lower ledges. The albite schist sequence of the Hoosac Formation and the amphibolites of the Turkey Mountain Member of the Hoosac are exposed in the upper ledges and constitute an outlier of Cambrian rocks surrounded by Wilmington Gneiss. Return to the cars and proceed north.

3.60 On the west side of the road, outcrops of Wilmington Gneiss dip gently to the northwest at 15°. There is a strong quartz rodding and biotite lineation nearly down the dip.

Proceed north to the end of Boyd Hill Road.

5.20 Turn right and proceed to Wilmington Center (0.45 mile).

Mileage (cont'd)

5.65 At the lights in Wilmington Center, turn right (east) on Route 9.

6.35 On the hill to the north are rusty weathering, albite schists dipping gently to the north toward the center of the Wilmington syncline.

6.80 Coombs' Sugar House where the field trip began. Latecomers may join the trip at this time (approximately 10:15 a.m.) and place. Continue east on Route 9.

7.10 Turn left at the NAJEROG sign and proceed 0.4 mile along an unpaved road (the original Molly Stark Trail).

Stop 2a. TURKEY MOUNTAIN MEMBER

Park at a white house and barn belonging to Mr. and Mrs. Donald Koelsch. Go up the hill to the west, observing the Turkey Mountain Amphibolite Member and the overlying sequence of albite schists of the Hoosac Formation (Modes of the Hoosac Formation are in Tables 15-19, p. 68 Skehan, 1961). Proceed to the top of the hill where you may observe recumbent folds in the albite schist and the development of small-scale microcline pegmatites. Structural analysis indicates that the sense of movement of the upper beds is toward the east relative to the lower. Such folds have been described by Skehan (1961, p. 103) as cascade folds. They are commonly well displayed in association with the Sadawga Pond dome and to a lesser extent with the Lake Rayponda dome. Commonly throughout southern Vermont such minor folds are late structures and are selectively better developed on the eastern flanks of the domes. The fact that the cascade folds seem not to be equally well developed on all sides of the dome calls into question Skehan's (1961) and Thompson's (1950) earlier conclusion that these folds are a product of the doming. Such cascade folds as seen at Stop 1 may, however, be related antithetically in some as yet unexplained way to the development of nappe structures in the "upper decken."

Return to Route 9 and proceed east.

7.10 Albite schist of the Hoosac Formation on the left.

Mileage (cont'd)

7.60 Pass Shearer Hill Road on the right.

7.95 Thick-bedded albite schist (similar to those rocks referred to by Hitchcock *et al*., 1861, as the gneiss at Jacksonville).

8.45 Junction with road leading north to Lake Rayponda.

9.40 Stop 3. HOOSAC FORMATION

Park on the west side of the road at the picnic and rest area. Outcrops of fairly massive, slightly schistose, nearly vertical beds on the east side of the Molly Stark Trail. These vertically dipping beds strike N.55°E. and consist of garnet-biotite-muscovite-quartz-albite schist alternating with less micaceous, more gneissic beds and thin quartzites. The albite schist encloses buff-weathering calcite lenses and pods 1/4-2 in.

This is one of only two localities in the Hoosac Formation where carbonate lenses or beds have been observed in the Wilmington-Woodford area. If the Hoosac Formation proves to be a facies equivalent of the Readsboro albite schist, these carbonate pods may be the easternmost exposures of albite schists enclosing the Sherman Marble Member of the Readsboro Formation.

Continue east on Route 9.

10.30 Stop 4. CHESTER AMPHIBOLITE

Park on the roadside near Hogback Ski Area. Observe the dominantly easterly-dipping folded beds of the Chester Amphibolite near the Skyline Lodge and Restaurant. The Chester Amphibolite here is characteristically a well-laminated ankerite-bearing epidote-chlorite-hornblende schist with quartz lenses. Note overturned synformal fold plunging to the northeast in which the axial plane dips north. These beds and exposures of the intensely folded Chester Amphibolite and the Stowe Formation, well exposed for the next 3/4 mile on Route 9 east, are near the axis of the Hogback syncline. Figures 20-23 in Skehan (1961) are photos taken at Skyline and Hogback Mountain. On a clear day Shelburne Mountain and the Holyoke Range, Massachusetts may be seen from this stop along with Mount Monadnock and the White Mountains of New Hampshire.

Mileage (cont'd)

Some may wish to walk east along Route 9 approximately 1.1 miles to Stop 5. The drivers and those wishing to ride should proceed east on Route 9, a distance of 1.3 miles.

11.60 At the junction of the road to Adams School, Marlboro, turn right and park along the unpaved road leading south. Walk back along the highway 0.2 mile, and proceed westerly along a logging road to an ankeritic steatite deposit near the boundary of the Stowe and Moretown Formations (Fig. 1).

Return to cars. A trip to alternate Stop 5a can be made by continuing east on Route 9 as follows:

4.75 Boudinaged amphibolite of the Stowe Formation on the north side of the highway.

5.50 Stop 5a. VIEW STOP

Just beyond the crest of a steep hill with large outcrops on either side of the highway, take a sharp right turn into the driveway of the Golden Eagle Motel. Park out of the way in the driveway. The view to the north looks toward Central Mountain on the boundary of the Wilmington and Brattleboro quadrangles, which is underlain by the rocks of the Missisquoi Formation (Fig. 1). This view is shown in Skehan (1961, Fig. 26, p. 89). Proceed westerly on foot along Route 9 for 0.2 mile. The first outcrops on either side of the highway are thin-bedded amphibolites at the top of the Stowe Formation. More ample descriptions of the rocks of the Stowe Formation are given on pages 85-88 of Skehan (1961). Mapping by Osberg (1965) has led to a number of correlations that indicate that the Stowe Formation is probably partially Cambrian and partially Ordovician in age.

Within the Wilmington-Woodford area only two outcrops of unmetamorphosed basalt have been observed, one of which cuts the amphibolites of the Stowe Formation at this locality and may be observed on the south side of Route 9. The other is at Stop 3 of the Sunday Field trip of Skehan (this Guidebook). Outcrops of very coarse-grained, flattened and rotated garnets may be observed on the north side of the road on the Tannelli property. Please do not engage in unrestrained collecting of specimens of these rocks.

Mileage (cont'd)

Return to cars and proceed west, returning to the intersection where cars were parked for Stop 5. Go south 0.55 mile on the road to Adams School at the crossroads. Continue south to Jenckes' Farm.

13.5 Stop 6. PINNEY HOLLOW FORMATION

Turn into Jenckes' Farm and park at the top of the driveway. Proceed south to the artificial pond and begin traverse downstream in the Green River. Excellent exposures of the Pinney Hollow Formation. A mode is presented in Skehan (1961, Table 21, p. 78) for a garnet-chlorite-biotite-muscovite-quartz schist from the Pinney Hollow Formation. Table 30 (p. 133) gives the chemical analysis of the chlorite from this formation. A traverse will be made downstream and up-section through the green schists and minor amphibolites of the Pinney Hollow Formation. The Pinney Hollow at this locality is typical of the Pinney Hollow of the east flank of the Green Mountains. For those making the Sunday Trip with Skehan (this Guidebook) note this rock and compare it with rocks to be seen at Sunday Stop 7.

The Chester Amphibolite is well-exposed in the Green River and here is overlain by the thin sequence of black schists and quartzites of the Ottauquechee Formation and the garnetiferous chlorite-muscovite-quartz schists and minor amphibolites of the Stowe Formation. The Stowe Formation is lithologically indistinguishable in hand specimen or thin section from Pinney Hollow and from the green schists of the Cavendish, except that the latter are commonly more intensely deformed. In certain localities, the Pinney Hollow, the Stowe and the Cavendish carry chloritoid.

Return to cars and continue south along road.

14.60 Junction Green River Road, turn left (east).

14.70 Junction with road to West Halifax, continue straight (east).

14.90 Stop 7. MORETOWN MEMBER, MISSISQUOI FORMATION

Park on the side of road. Outcrops in field to north of road are the Moretown Member of the Missisquoi Formation. Rocks are fairly typical, light gray quartz-mica schists with interlaminations of quartz-rich granulite and mica schist. The scale of the interlamination ranges from 0.5 mm. to several centimeters. Porphyroblasts include garnet, chlorite, and biotite.

Mileage (cont'd)

Return to cars, continue east on Green River Road.

15.30 Entering Brattleboro quadrangle.

15.60 Road junction in Harrisville, continue straight (east) on the Green River Road.

16.00 Stop 8. BARNARD VOLCANIC MEMBER, MISSISQUOI FORMATION

Park cars on the side of road. Small outcrops along the north side of the road are in the Barnard Volcanic Member. These outcrops show some of the variety of the rocks in the Barnard. Hornblende-plagioclase amphibolite, epidote amphibolite, and porphyritic amphibolite with prominent feldspar megacrysts are exposed here. Dark gray to brown weathering chloritic schists and light-colored felsic gneisses and schists are not as abundant here as they are elsewhere in the member.

Continue east on Green River Road.

16.20 Outcrops of Cram Hill Member, Missisquoi Formation.

16.50 Stop 9. CRAM HILL MEMBER, MISSISQUOI FORMATION

Park on the side of the road. This outcrop is typical of the Cram Hill Member, which consists largely of rusty-brown weathering, fine-grained, black phyllite and schist. The rusty weathering is due to finely disseminated pyrite and pyrrhotite. The only porphyroblasts large enough to be seen here are biotite and pyrite. Bedding is difficult to distinguish where thin black quartzite interbeds are not present. A few amphibolites are present in the outcrop. A secondary cleavage cuts the schistosity here, causing the rock to break into elongated tabular blocks used locally as fence posts.

17.50 Crossing Middle Ordovician-Silurian unconformity (not well-exposed along the road).

17.60 Outcrops of Northfield Formation.

18.60 Stop 10. NORTHFIELD FORMATION

Park with care along side of road. The Northfield Formation is a gray quartz-muscovite schist with conspicuous garnet porphyroblasts 1 to 2 mm. in diameter. Biotite porphyroblasts are also common. The dip direction of the principal schistosity is changing in this area from

Mileage (cont'd)

east-dipping off the Green Mountain anticlinorium to west-dipping adjacent to the Guilford dome to the east. A prominent slip-cleavage is present here and is probably related to the rise of the Guilford dome. The slip-cleavage is developed parallel to the axial surfaces of the minor folds of the third stage. The ubiquitous crinkles are the result of the intersection of the schistosity and slip-cleavage surfaces.

19.50 Stop 11. WAITS RIVER FORMATION

Park at the bottom of the hill and walk back uphill to outcrops on the north side of the road. These are fairly typical of the more calcareous portions of the Waits River Formation. The impure marbles weather a punky-brown with a friable surficial rind of the non-carbonate minerals left by the leaching of the carbonates. The fresh marble is steel gray. The modal percentages of carbonates in the impure marble beds range from 35 to 70 percent. Quartz accounts for most of the rest of these beds with minor amounts of muscovite, biotite, garnet, plagioclase and actinolite present. Note the small "skarn" reaction zones at the contact of the marble beds and the surrounding mica schists. The large folds in the marble beds seen here are tentatively correlated with the second stage of minor folding, that congruous with the development of the Prospect Hill recumbent fold.

Return to cars and continue east on Green River Road.

19.60 Guilford-Halifax town line.

20.20 Outcrops of Waits River Formation to north.

20.70 Road junction, turn right (south) toward village of Green River.

21.30 Stop 12. WAITS RIVER, STANDING POND, AND GILE MOUNTAIN FORMATIONS

Park near abandoned farm house. Walk to north end of pasture. From here walk southwest across the pasture through the units in the Prospect Hill recumbent fold. The recumbent fold has been arched by the later doming so that now the units at this locality dip southwest away from the axial trace of the Guilford dome. The exposures in the pasture and along the base of the hill show a nearly continuous section through the Standing

Mileage (cont'd)

Pond and Gile Mountain Formations. At the north end of the pasture, observe the contact of the Waits River Formation with the amphibolites of the Standing Pond. Note garnets to 1/2 inch but please do not remove them. Cross the Standing Pond amphibolites (small ridges) and interbedded brown weathering schists (small gullies). On the second small ridge note the contact of the amphibolite with a schist interbed. Which way is up? The Standing Pond here is some 400 feet thick. The Gile Mountain is exposed along the base of the hill and along the road (see as follows).

The Gile Mountain Formation consists of feldspathic and micaceous quartzites with thin mica schist interbeds. Kyanite and staurolite occur locally in these interbeds. The contact of the Gile Mountain with the Standing Pond at the southwest side of the pasture is placed at the appearance of the first amphibolite. A few thin beds of impure quartzite similar to those in the Gile Mountain occur in the Standing Pond here. Note the development of thin pink bands of coticule (spessartine garnet and quartz) in the Standing Pond near the contact with the Gile Mountain in the woods at the southwest end of the pasture.

Return to road and to cars.

Stop 12a.

Walk south along road 0.1 mile to road cut in the typical interbedded feldspathic and micaceous quartzites and mica schists of the Gile Mountain Formation. The quartzite beds range from a few inches to 3 feet thick.

END OF FIELD TRIP

Return to cars, turn around, head north and proceed to Brattleboro and Burlington.

22.00 Junction with road to Brattleboro, turn right (east).

22.50 Outcrop of Waits River Formation dipping west off the Guilford dome. (Note hill to west--Governors Mountain--formed by the west-dipping Gile Mountain Formation and Standing Pond Volcanics in the Prospect Hill recumbent fold. Slopes on the west side of hill are essentially dip slopes.)

Mileage (cont'd)

23.40 Outcrop of Waits River Formation.

25.60 West-dipping Waits River Formation.

27.10 Junction with unpaved road to left, bear right (stay on paved road).

28.60 Junction Route 9 in West Brattleboro. Turn right (east) and continue approximately 2.5 miles to junction Interstate 91. Take 91 north to I-89 to Burlington. (Note--In the northbound entrance to I-91 from Route 9 there are excellent exposures that include the eastern band of the Standing Pond Volcanics. Here the Standing Pond is a quartz-plagioclase-hornblende-biotite granulite due to the lower, biotite zone, metamorphic grade.)

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Trip B-2

MAJOR STRUCTURAL FEATURES OF THE TACONIC ALLOCHTHON IN THE HOOSICK FALLS AREA, NEW YORK-VERMONT

by

Donald B. Potter
Hamilton College

<u>Purpose:</u> We will see on this trip the two major thrust sheets that comprise the eastern part of the Taconic allochthon in this area. We will examine in detail some of the thrust contacts, and see the recumbently folded nature of the base of the lower thrust sheet. We will also see the Middle Ordovician submarine slide breccia, with its giant clasts, that occurs immediately beneath the allochthon.

<u>Background and acknowledgements</u> This trip is based on a ten-year detailed stratigraphic and structural study (Potter, 1972) in the Hoosick Falls area (Figure 1). Field work has been supported by the New York State Geological Survey, the National Science Foundation, The Geological Society of America, and Hamilton College. Lane (1970) has made a detailed structural analysis at selected localities in the area aimed at deciphering the deformational history. His work is not intended to be an assessment of thrust-no thrust problem.

E. Zen, W. Berry, J. Bird, G. Theokritoff, and D. Fisher have greatly aided Potter's study through field visits, identification of fossils, and through published data (see Zen, 1967 and references cited therein.)

Prior to the present work the most definitive study in the Hoosick Falls area was by Prindle and Knopf (1932). Bonham (1950), Balk (1953), and Lochman (1956) have also contributed to our knowledge of the geology and paleontology of this area. MacFadyen (1956), and Hewitt (1961) mapped the quadrangles east and northeast, respectively, of the Hoosick Falls area; and Metz (1969) has recently mapped the Cambridge Quadrangle to the north.

<u>Stratigraphy</u> While not the major concern of this trip, the stratigraphy of this area must be understood in at least summary fashion for the stratigraphic details enable us to establish structures which constitute prime evidence for the major thrusts. Figure 2 summarizes the relations of the two major stratigraphic sequences.

The Taconic Sequence, comprising the allochthon, is approximately 4000 feet thick, and consists of turbidites and pelites suggesting deposition in deep water with unstable bottom conditions: delicately laminated argillite and thin-bedded chert suggest deep, quiet water conditions; euxinic conditions are suggested by pyritiferous black slate with and without graptolities; transportation and deposition by turbidity currents is indicated by the litholigic character, graded bedding and sole markings of the major graywacke units; unstable bottom conditions and submarine slumping are indicated by intraformational breccias (ibc, Figure 2) and by the presence of a few exotic clasts in some of the units. Stratigraphic units within the Taconic Sequence show great continuity north and south within the allochthon, but exhibit maximum change in thickness and in lithic character east-west (across strike). Thus, practically every unit shown in Figure 2 can also be identified 60 miles

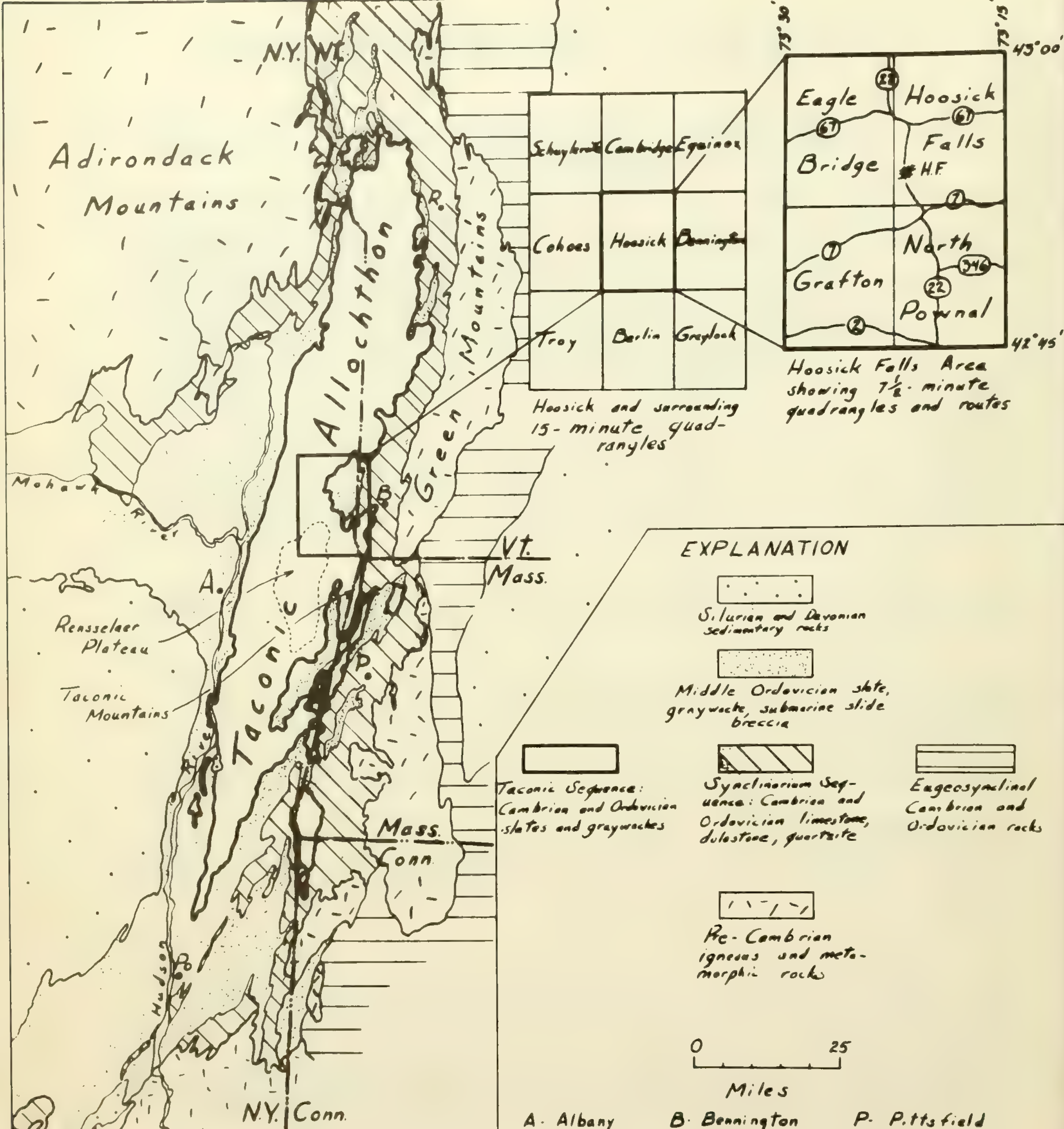


Figure 1. Geologic setting of the Taconic Allochthon and location of the Hoosick Falls Area

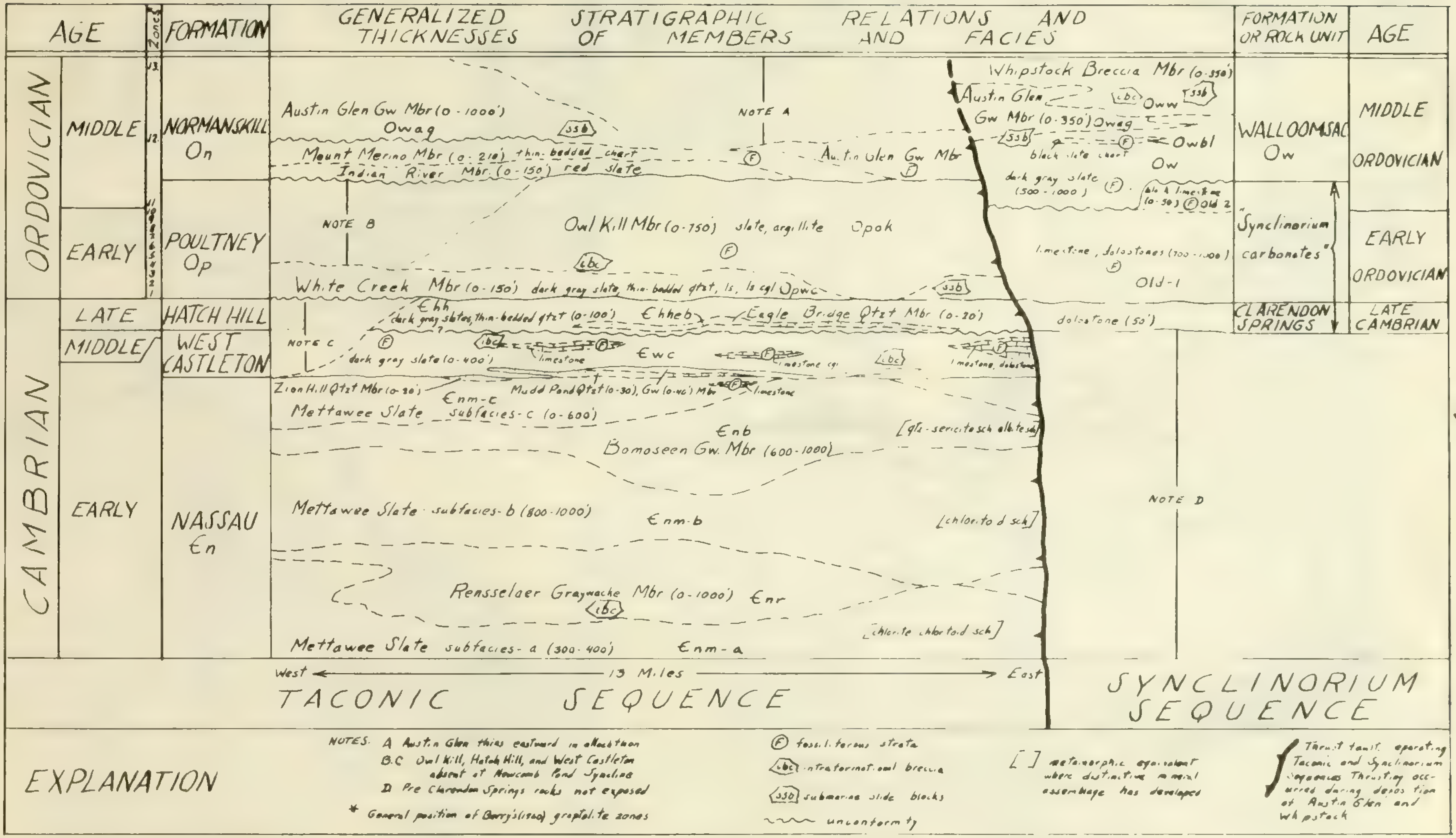
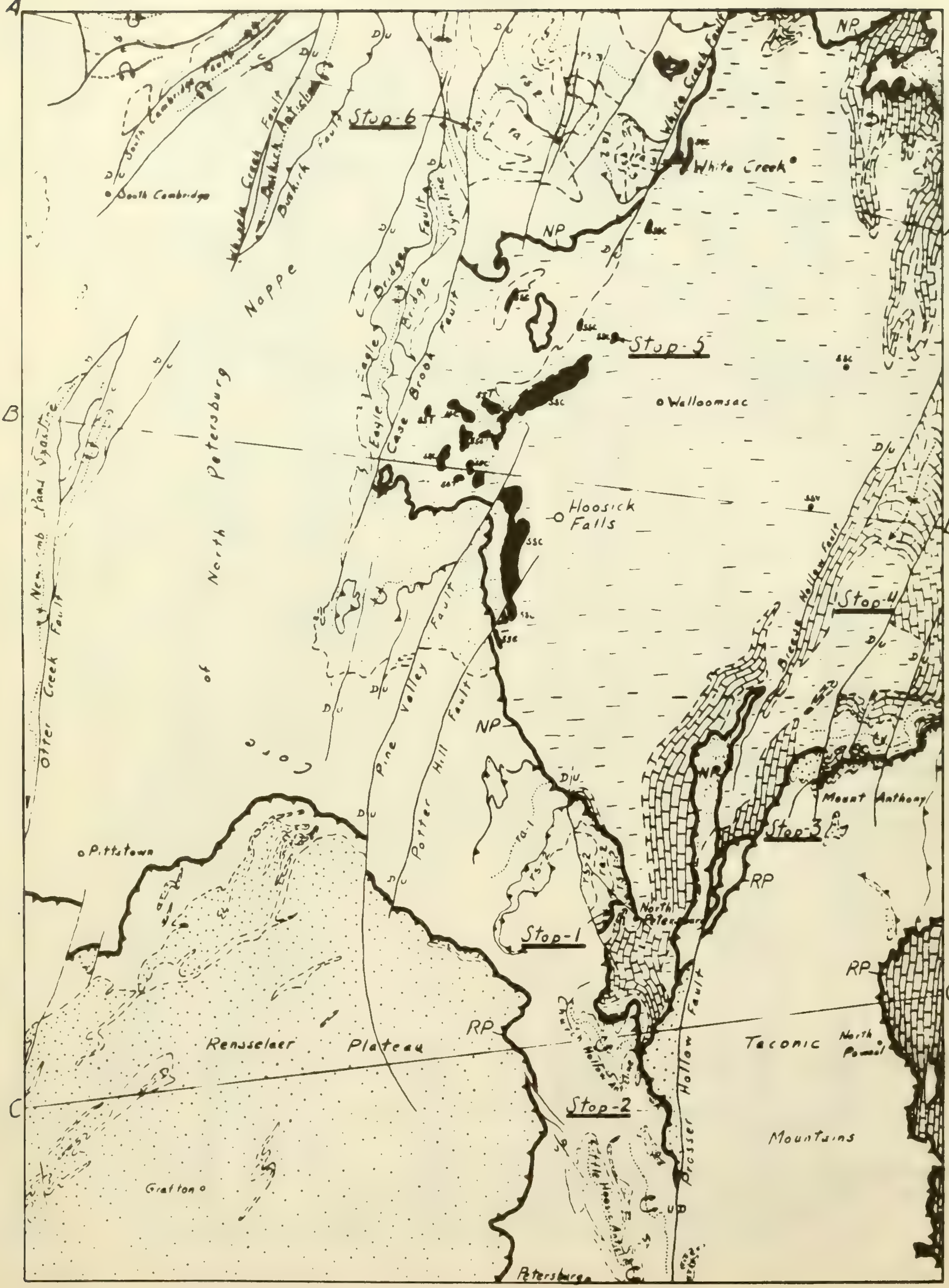


Figure 2. Stratigraphy of the Hoosick Falls Area


A
A'
B
B'
C
C'
Stop-1
Stop-2
Stop-3
Stop-4
Stop-5
Stop-6
White Creek
White Creek Fault
South Cambridge
South Cambridge Fault
Whelk Creek Fault
Batten Kill Anticlinorium
Bunkirk Fault
Eagle Bridge Fault
Eagle Bridge Syncline
Case Brook Fault
Walloomsac
Hoosick Falls
Core of North Petersburg Nappe
Newcomb Pond Syncline
Otter Creek Fault
Pine Valley Fault
Potter Hill Fault
Breese Hollow Fault
Mount Anthony
North Petersburg
Pittstown
Rensselaer Plateau
Taconic Mountains
North Pownal
Grafton
Petersburg
Brosser Hollow Fault
NP
RP
D/U

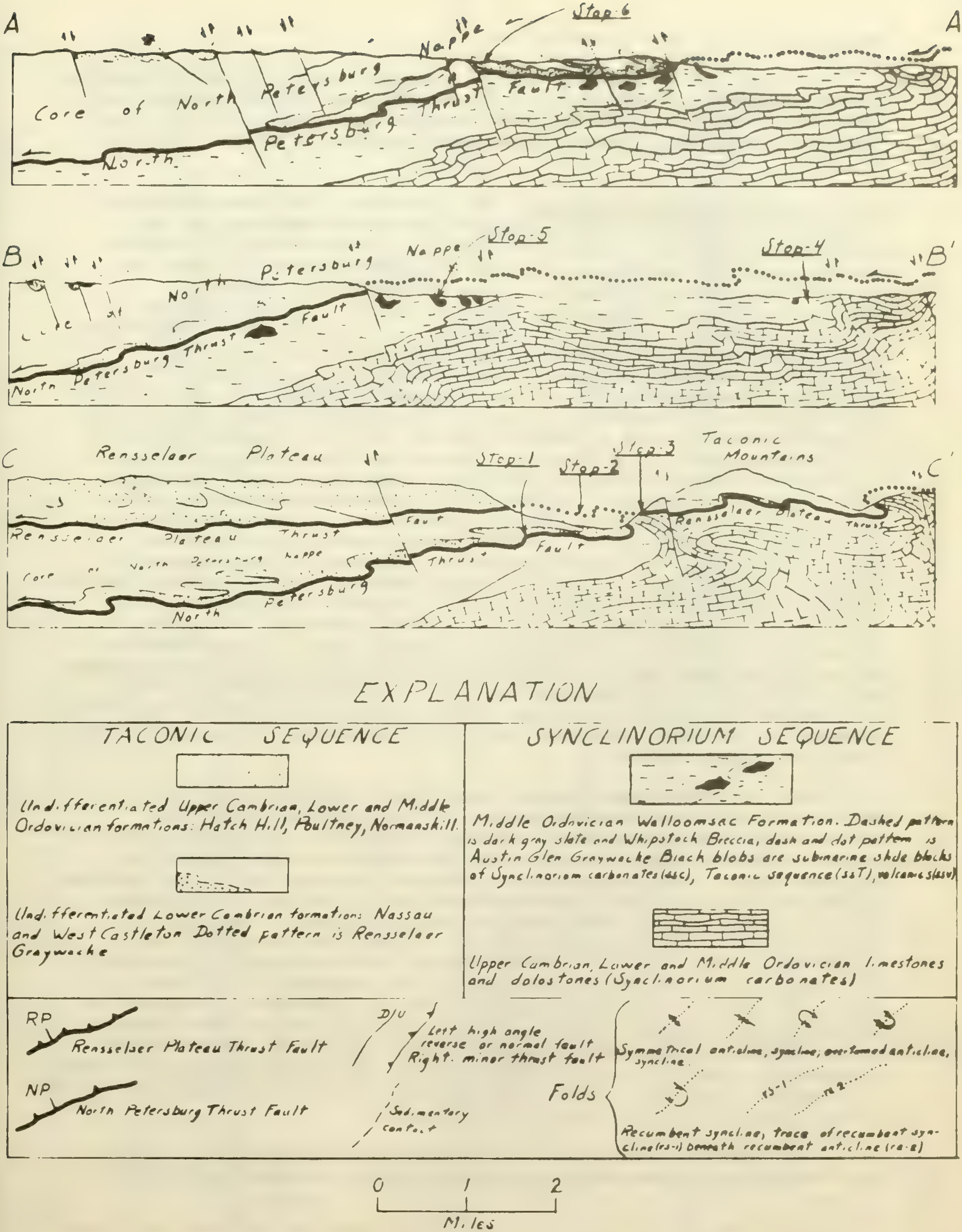


Figure 3. Geologic Map and Structure Sections of Hoosic Falls area.

north, at the north end of the allochthon.

The Synclinorium Sequence, largely synchronous with the Taconic Sequence, is about 2000 feet thick and consists of limestones, dolostones, and quartzites that attest to a shallow water shelf environment. These are overlain by slate-graywacke-submarine slide breccia that mark the period of Middle Ordovician thrusting.

Major structures and their evolution The allochthon in this area consists of two sheets, one above the other, that have been thrust westward onto the Synclinorium Sequence. Evidence for thrusting includes lithologic contrasts and gross structural discordance between synchronous formations above and below the thrust traces (Figure 3); slices of carbonate rock from the Synclinorium Sequence between the two thrust sheets (Stop 2); crushing, shearing, and mineralization at the thrust zones. The lower (North Petersburg) thrust sheet includes all the rocks of the Taconic Sequence except the Rensselaer Graywacke; recumbent folds are extensively developed in the lower 1000 feet of this sheet which consists of younger formations than the upper part of the sheet (structure sections, Figure 3). The North Petersburg sheet is thus a huge recumbent anticline or nappe (Figure 4), and it is correlated with Zen's (1967) Giddings Brook slice (Figure 5). Beneath the North Petersburg thrust is the Middle Ordovician Walloomsac formation consisting of slate, graywacke, and submarine slide breccia. The Whipstock submarine slide breccia contains clasts of the Taconic Sequence and some giant blocks of carbonate rocks from the Synclinorium Sequence. It is inferred that thrusting was a submarine phenomenon, that Austin Glen Graywacke was deposited on both Taconic and Synclinorium sequences at the early stages of orogeny, that as the thrust sheets moved into this area from the east, blocks of limestone and dolostone up to 1.8 miles long and 700 feet thick (from the shelf environment) and blocks of Taconic Sequence rocks (from the advancing thrust sheets) slid westward into the mud in the deeper parts of the basin to form the Whipstock. Unconsolidated breccia, graywacke, and mud were overridden by the North Petersburg sheet, and, because of the gross overturning of this sheet, unconsolidated Austin Glen Graywacke of the Taconic Sequence was locally melded with the unconsolidated material beneath the thrust.

The upper (Rensselaer Plateau) thrust sheet is perhaps the eastern core of the North Petersburg nappe which was thrust westward onto the core and inverted limb of the leading part of the nappe (Figure 4). On the plateau the Rensselaer Plateau sheet consists of Rensselaer graywacke and underlying Mettawee slate. Eight formations or stratigraphic units of the Taconic Sequence, ranging from Early Cambrian to Middle Ordovician, constitute the Rensselaer Plateau sheet on Mount Anthony and the Taconic Mountains. Identification of these units rules out MacFadyen's (1956) conclusion that the schists and related rocks here, which he called the "Mount Anthony Formation," are Middle to Upper (?) Ordovician and autochthonous. The correlation of this thrust sheet here with that capping the Rensselaer Plateau is based on the extensive exposures of Rensselaer Graywacke at the base of the sheet on Mount Anthony and on the Taconic Mountains (Figure 3), and on the fact that no other thrust sheet occurs between this one and the North Petersburg sheet or the autochthonous rocks below. Thus, Zen's (1967) Dorset Mountain slice in this area is considered to be the Rensselaer Plateau thrust sheet (Figure 5).

Both major thrust planes and thrust sheets have been refolded by a later stage of deformation that produced a pervasive slaty cleavage-foliation. All

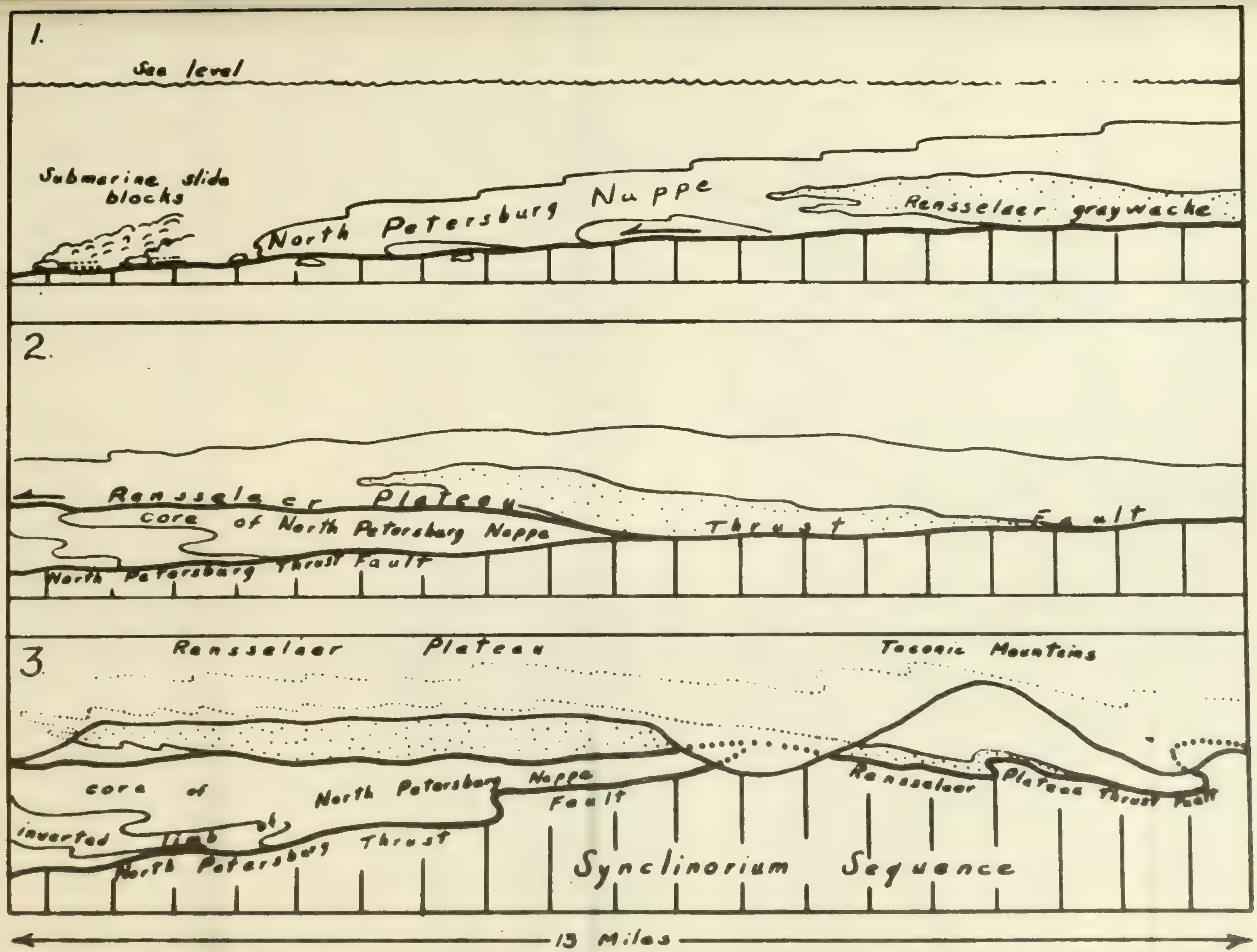


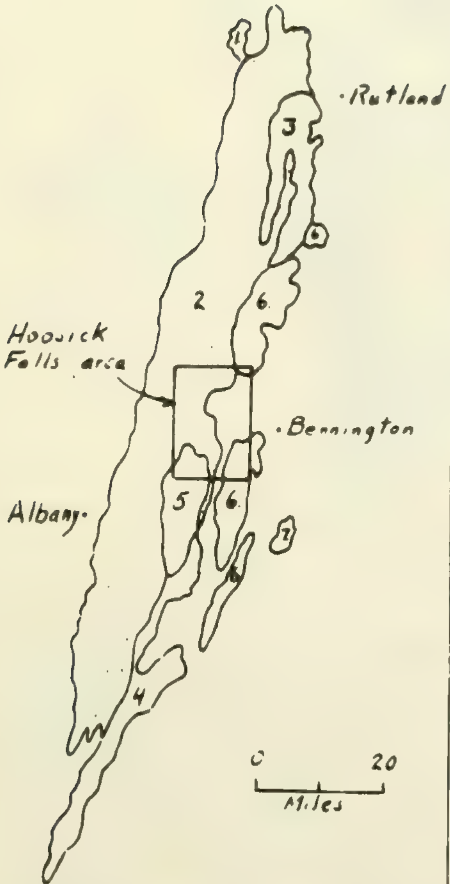
Figure 4. Schematic sections showing emplacement of thrust sheets

the rocks in the area underwent a regional metamorphism in Middle Ordovician time. Increase in rank from west to east is shown by the recrystallization of limestones and dolostones, and by metamorphism of argillites and slates to phyllites and schists containing chlorite, chloritoid, sericite, and albite. High-angle reverse and normal faults, striking north-northeast, cut the two thrust sheets and the autochthonous rocks beneath.

Lane's work suggests that four deformational episodes can be recognized in this area. The first, D_0, occurred at least in part before complete lithification of sediments, and consisted of large-scale westward transport and formation of recumbent folds and nappe structures. The next episode, D_1, produced a system of NNE-trending, westward-overturned folds and a pervasive axial plane slaty cleavage, S_1. Extensive mylonite zones along the Rensselaer Plateau thrust were formed contemporaneous with S_1, and metamorphism also occurred at this time. After the formation of S_1 and the mylonites, minor movement occurred on the Rensselaer Plateau thrust and perhaps on the North Petersburg thrust as well. D_0 and D_1 established the overall geometry of structures in the central Taconics. Later deformations, in this area at least, served only to modify the structure. During D_2, the pervasive slaty cleavage was refolded on a NNE axis, and an axial plane slip cleavage, S_2, was locally developed. The final episode of deformation, D_3, caused folding of S_1 about an ESE axis, and locally developed axial plane slip cleavage, S_3. The use of the term "deformational episode" is not intended to imply knowledge of temporally discrete deformations. It is possible that some of the deformations described may have been essentially continuous.

STOP DESCRIPTIONS

The location of each stop is plotted on the geologic map, Figure 3, and the general structural setting at each stop is indicated on the structure sections. The topographic maps (1:24,000) accompanying descriptions of Stops 1-6 show the limit of outcrops (fine dotted lines) and main geologic contacts. Refer to Figure 2 for letter symbols of stratigraphic units.



Thrust slices in Taconic Allochthon (abridged from Zen, 1967, Fig. 2)

Zen's (1967) thrust slices (in order of relative age of emplacement, oldest at bottom)	Thrust sheets in the Hoosick Falls Area (this report)
6 Dorset Mtn and Greylock slices	Rensselaer Plateau sheet
5 Rensselaer Plateau slice	
4 Chatham slice	
3 Bird Mountain slice	
2 Giddings Brook slice	North Petersburg sheet
1 Sunset Lake slice	
	NOTE: Zen's slice 6 in southeastern part of Hoosick Falls area is interpreted to be eastern half of Rensselaer Plateau sheet; in northeast corner of area, slice 6 is mapped as North Petersburg sheet.

Figure 5. Correlation of North Petersburg and Rensselaer Plateau thrust sheets with Zen's (1967) slices.

STOP-1

North Pownal
Quadrangle

0 1000
Feet

Exposure of North Petersburg thrust fault, west of North Petersburg. Walk to thrust contact on steep slope (elevation 800 feet) via outcrops of limestone and dolostone of the Synclinorium carbonates. Steep slope above carbonates and below thrust is believed to be largely underlain by Walloomsac slate. There are a few small outcrops of slate on this slope, and a good exposure of the slate beneath the thrust fault 3/4 mile north of this stop. Less than five feet beneath the thrust fault, and inbedded in Walloomsac slate, is a large block of limestone (Synclinorium carbonate), interpreted to be a submarine slide block.

The thrust zone is characterized by shearing, mylonitization, bleaching and calcification of argillites, cherty argillites, and slates belonging to the Owl Kill Member of the Poultney Formation. Above this, through a vertical distance of some 200 feet, is an inverted sequence of the White Creek Member of the Poultney (ribbon limestones in black slate), Hatch Hill (thin-bedded quartzites in dark gray slate), Eagle Bridge Quartzite, and Bomoseen Graywacke. The Bomoseen marks the core of recumbent anticline, ra-2 (Figure 3), one of several recumbencies in this part of the area that characterize the lower part of the North Petersburg nappe.

STOP-2

North Pownal
Quadrangle

0 1000
Feet

Exposure of the Rensselaer Plateau thrust fault north of Prosser Hollow. Below the thrust fault is an apparently normal sequence of Bomoseen, Mettawee, and Hatch Hill (with Eagle Bridge Quartzite) - all part of the North Petersburg thrust sheet. The Rensselaer Plateau thrust fault is marked by a large sliver of limestone and dolostone of the Synclinorium carbonates that have been tectonically dragged to their present position. Immediately above the Rensselaer Plateau thrust is the Rensselaer Graywacke, perhaps several hundreds of feet thick and intensely sheared. The graywacke is faulted against chloritoid schist (Mettawee) 0.3 miles east of this stop.

The following details of the fault zone are noted. First, the Rensselaer Graywacke above the thrust is mylonitic through a zone approximately 150 feet thick (measured perpendicular to foliation), and the mylonitic foliation is concordant with normal foliation above and below the thrust zone (Figure 6). Second, the thrust plane truncates the mylonitic foliation. Third, a well-developed foliation parallel to the thrust plane occurs in the uppermost 2-3 feet of the limestone. Numerous other structural features may be observed. Widely spaced fractures parallel to the thrust plane also truncate the foliation and show a similar sense of movement to that on the thrust. Several warps in the thrust plane apparently represent areas where (later) movement on the thrust has locally followed the foliation instead of cutting across it. Near the upper (western) end of the outcrop, a sliver of mylonitic graywacke about 5' x 5' is completely enclosed within the limestone. West of this, the thrust plane steepens and follows the trend of the foliation in the graywacke for an indefinite distance.

Two generations of folds are occasionally visible in the mylonites above the thrust. In one generation, the axial planes are parallel to the foliation; the axes generally trend to the north but are variable. In some cases, the plunge of the axes is perpendicular to the strike of the axial plane, thus forming a reclined fold. This fold style is common in other thrust zones, notably along the Moine thrust in the Scottish Highlands. The second visible generation of folds has NNE trending axes and nearly vertical axial planes. These folds are correlated with F_2, one of the four fold systems in the non-mylonitic rocks in this area.

Interpretation: The earliest structural event well-represented at this stop is the formation of the pervasive axial plane foliation, S_1, and the accompanying regional metamorphism. Emplacement of the graywacke and chloritoid schist along the Rensselaer Plateau Thrust may have occurred prior to the formation of S_1. Evidence for this is the occurrence in several places along the thrust of tectonic slivers of autochthonous carbonates around which S_1 has been refracted.

The mylonites either were pre S_1 and rotated into their present orientation during the formation of S_1, or else formed at the same time as the foliation. Following the ideas of Johnson (1967) the latter explanation is preferred. The mylonites are not necessarily related to large scale thrust movement, and consequently, evidence for emplacement of the Rensselaer Plateau thrust must come mainly from regional stratigraphic and structural studies.

Following S_1, minor movement occurred between the graywacke and the slates beneath. This movement caused the presently observed thrust plane, the thin zone of well-developed foliation in the upper few feet of the limestone, and the low angle fractures in the rocks immediately above and below the thrust. In other localities, notably west of the Little Hoosic Valley, and at Stop 3, the later movement caused a marked, but local, disturbance of foliation.

<u>Explanation of Figure 6.</u>

Equal area diagrams (lower hemisphere) showing orientation of poles to foliation in vicinity of Rensselaer Plateau thrust at Stops 2 (A-C) and 3 (D-F). Contours are 15%, 10% and 5% per 1% area unless otherwise indicated.

A. Below thrust at STOP-2: 105 measurements.

B. Thrust zone, within 5 feet of thrust plane, at STOP-2: 28 measurements.

C. Above thrust at STOP-2: 42 measurements.

D. Thrust zone, within 5 feet of thrust plane at STOP-3: 44 measurements.

E. Above thrust, 20 to 50 feet vertically up slope to the NE of STOP-3: 20 measurements.

F. Regional trend of foliation in NE 4/9 of North Pownal Quadrangle. 410 measurements. Contours are 10%, 5% per 1% area.

These diagrams are intended only as a qualitative guide to foliation orientation. The contours are not statistically rigorous.

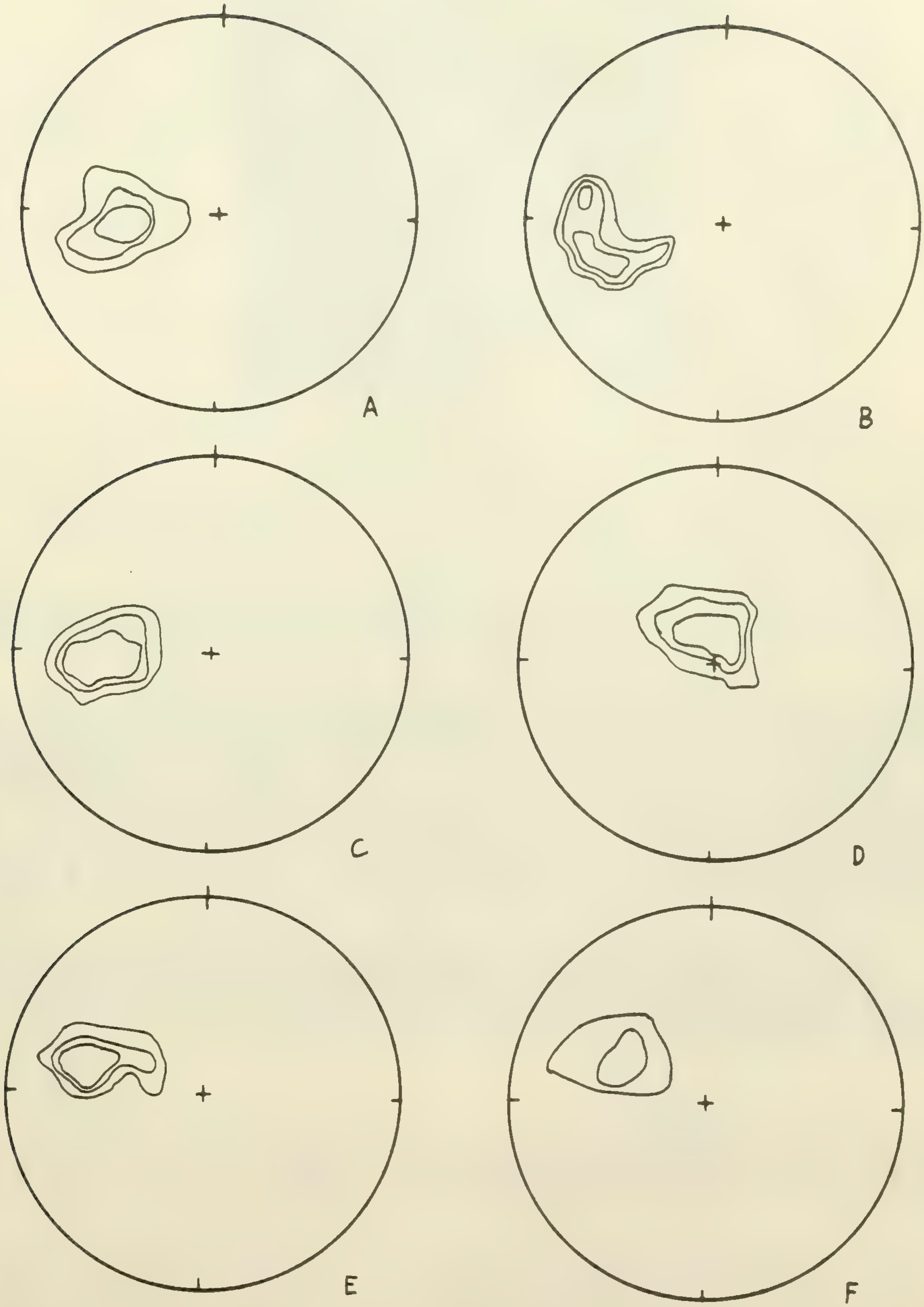
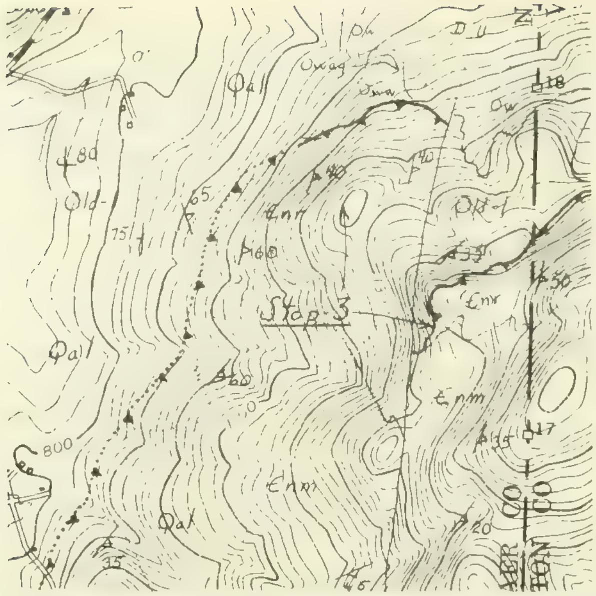


Figure 6 Equal area diagrams of poles to foliation in vicinity of Rensselaer Plateau thrust fault

STOP-3



North Pownal Quadrangle

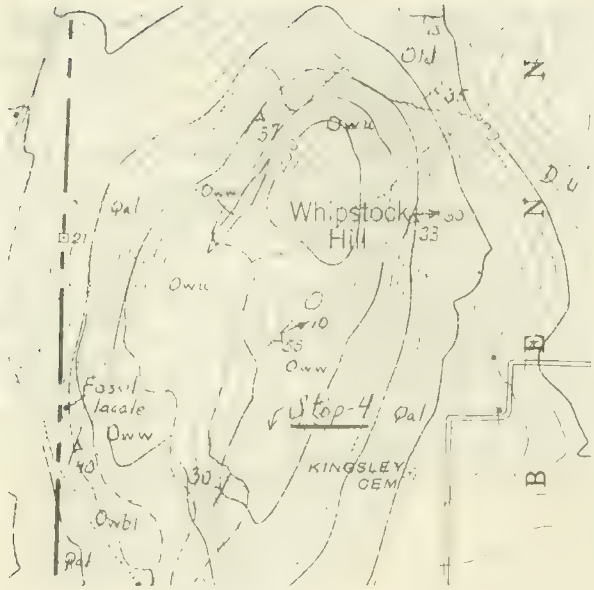
0 1000
Feet

Exposure of folded Rensselaer Plateau thrust fault on west shoulder of Mount Anthony (this stop treats you to a 3/4 mile ride or walk, 700 feet relief-one way). The rocks beneath the thrust are autochthonous Synclinorium carbonates-limestones and dolostones. In the low ground north of the shoulder of Mount Anthony these carbonates are recumbently folded (axis of fold trends east-west) with Walloomsac slate.

At closed contour 1300 we will see some least metamorphosed Rensselaer Graywacke. The thrust fault is exposed at elevation 1500. Immediately beneath the folded thrust plane the upper few feet of limestone is conspicuously thinly foliated, with foliation parallel to thrust plane. The graywacke above the thrust is schistose and contains characteristic seams of granular quartz. The foliation in the overlying graywacke is also parallel to the thrust plane, but within approximately 20-30' vertically, S_1 strikes NNE and dips moderately to steeply ESE, ie., concordant with the regional attitude of foliation (Figure 6). The graywacke is mylonitic for several hundred feet up the slope. In thin section and occasionally in outcrop, a slip cleavage is seen to displace the mylonitic foliation. This cleavage is not related to any of the regional fold systems.

As in Stop 2, the graywacke is believed to have been emplaced prior to the formation of the pervasive slaty cleavage and regional metamorphism. The mylonite formed at the same time as S_1, and was locally deformed by later movement on the Rensselaer Plateau thrust.

STOP-4



Hoosick Falls
Quadrangle

0 1000
Feet

Exposure of Whipstock Breccia at Whipstock Hill. Several small scattered outcrops of Whipstock Breccia can be seen in the grassy terrain at the 1180 crest of Whipstock Hill and these afford a close examination of the dark gray silty argillite or slate matrix, intraformational clasts, and a few exotic submarine slide blocks.

The matrix of the breccia is irregularily cleaved because of the intraformational clasts which range from about 1 mm. to 5 cm. in maximum dimension. Many of these small clasts lie in the plane of foliation and are smeared out and elongated so that they define a prominent lineation. The dominant clasts are fine grained serictic siltstone, green-gray argillite, quartzite, and fine grained limestone. The similarities between these lithologies and thin laminae and layers in the Walloomsac slate suggests an intrabasinal origin for the clasts. In addition, however, there are in the Whipstock exotic blocks of the Taconic Sequence, the Synclinorium carbonates, and volcanic rocks. Two exotic blocks can be seen at Whipstock hill: one is a conglomeratic quartzite (Rensselaer Graywacke or Zion Hill Quartzite), about 16 inches in maximum dimension, consisting of subrounded grains of quartz, quartzite, and oligoclase in a matrix of fine grained quartz, sericite, and chlorite. A second exotic block, about 4 x 4 feet in plan, is inequigranular pyritic quartzite resembling the Mudd Pond Quartzite.

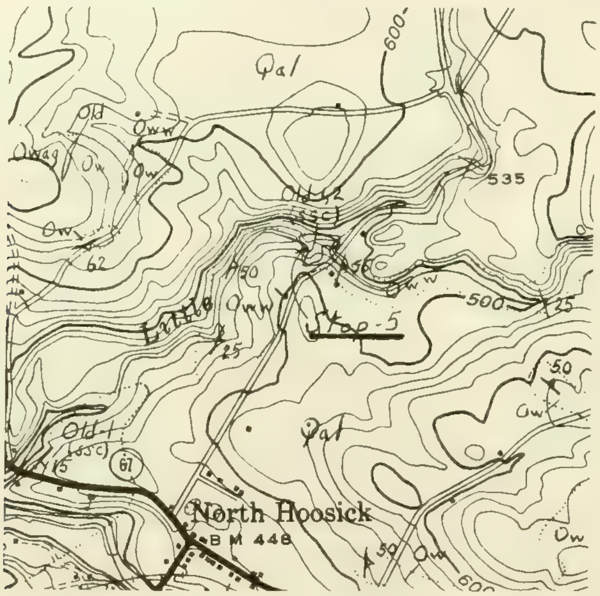
The Whipstock Breccia on Whipstock Hill is infolded with a large mass of fine grained phyllitic siltstone, and recrystallized radiolarian chert (Owu), rocks of unknown stratigraphic position. These rocks may be beds in the breccia, large intraformational clasts, or perhaps giant clasts of the Taconic Sequence.

The Whipstock is an integral part of the Walloomsac Formation. It is widely distributed beneath the North Petersburg thrust fault, and is closely associated with lenses of Austin Glen Graywacke. Its age is Wilderness or post-Wilderness for the breccia is underlain on the west slope of Whipstock Hill by a black slate containing graptolites of the <u>Climacograptus</u> <u>bicornis</u> Zone.

A similar submarine slide breccia (Forbes Hill) has been identified by Zen (1967) in the northern Taconics; and wildflysch conglomerates are extensively exposed at the west edge of the allochthon (see Bird, 1963, pp. 17-19).

The crest and west flank of the Green Mountains are visible northeast of Whipstock Hill. Pre-Cambrian gneisses of the Mt. Holly Complex are exposed along the crest of the range, and the Lower Cambrian Cheshire Quartzite, which rests unconformably on the Mt. Holly, forms prominent dip slopes on the mountain flank. The Cheshire here is at the base of the Synclinorium Sequence. The Taconic Allochthon, being intermediate in facies between that of East Vermont (eugeosynclinal) and the Synclinorium Sequence (miogeosynclinal) presumably came from an area east of the exposures of the Cheshire Quartzite. Zen (1967) has proposed the Green Mountain core as the likely root zone. The prominent hill to the southeast is Mount Anthony, and the base of its steep north face marks the trace of the Rensselaer Plateau thrust fault.

<u>STOP-5</u>



Hoosick Falls Quadrangle

0 1000

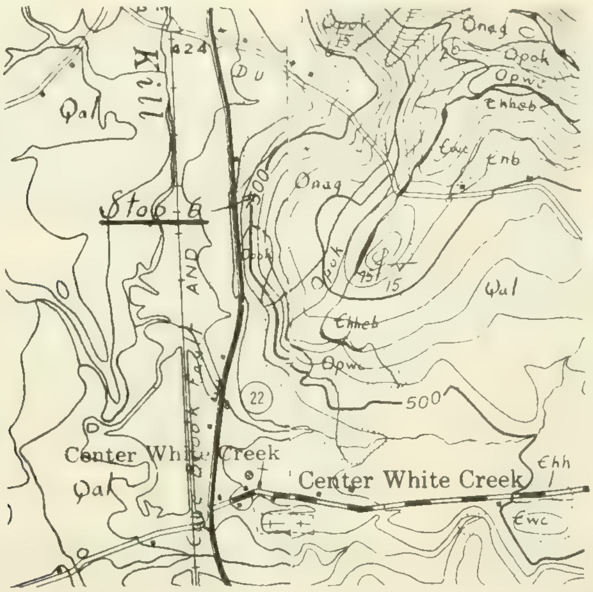
Feet

Exposure of large submarine slide block in bed of Little White Creek. The block here consists of Synclinorium carbonates and measures about 300 x 400 feet in plan. Whipstock Breccia is exposed above and below the block which consists of highly folded limestone and dolostone (Old-1) overlain by black argillaceous limestone (Old-2). The contact between the black limestone of the block and the overlying Whipstock is locally conformable, but at the west (downstream) end of the exposure the contact between layers in the block and foliation in the breccia is discordant, and the Whipstock Breccia at the contact is crumpled, sheared, and faulted. Slaty cleavage in the Whipstock near the block is cut by a younger slip cleavage.

The problem of identity of masses of Synclinorium carbonates as discrete submarine slide blocks ("ssb" on Figure 3) is difficult for exposures in the western part of the Hoosick Falls re-entrant are poor, and these carbonate masses could be isolated outcrops of the complexly folded carbonate sequence. The following evidence suggests a submarine slide block origin: 1. an apparent structural discordance occurs from one block to another in the area southwest of STOP 5; 2. discrete masses (blocks) of the Taconic Sequence

occur in the same terrain; 3. limestones at five of the "blocks" carry fossils of Early and Middle Ordovician age, yet the blocks are surrounded by Mid. Ordovician breccia, graywacke, or slate, or occur at the contact between graywacke and slate. The last relation suggests that submarine sliding occurred after a thick accumulation of mud (Walloomsac slate) and at the onset of deposition of the Austin Glen Graywacke.

STOP-6



Eagle Bridge and Hoosick Falls Quadrangles

0 1000

Feet

Recumbently folded Austin Glen Graywacke at roadcut on Route 22. The Austin Glen Graywacke, with interbedded dark gray slate and cross-bedded siltstones, is recumbently folded here in the core of recumbent syncline, rs-1, near the base of the North Petersburg thrust sheet. The exposure is apparently just east of the Case Brook reverse fault. Axial planes of recumbent folds here strike between N25° E and N80° E and dip from 13° to 23° southeast; axes plunge east-southeast from 3 to 20 degrees. Recumbent anticlines open to the northeast, synclines open to the southwest.

The exposure is notable for its wealth of primary sedimentary structures which includes subtle graded bedding and sole markings in the graywacke beds, and various types of cross-bedding in the siltstones.

MILEAGE LOG

Note: Depending on the size and interests of the group we may adhere strictly to the log or we may include unlogged localities to examine some of the stratigraphic and structural details within the Taconic and Synclinorium sequences.

Mileage	
00.0	Intersection of Routes 22 and 346 at North Petersburg, N.Y. This is the southern tip of the Hoosick Falls re-entrant. (See Figure 3) The valley bottom to the north and northeast is underlain by autochthonous carbonate rocks and slates of the Synclinorium Sequence. The North Petersburg Thrust Fault is exposed near the base of the steep hills to the west and northwest and above this thrust is the Taconic Sequence of formations. Turn north on Route 22
00.3	Park cars at transformer on west side of road and walk up steep slate-mantled slope, crossing a few outcrops of Synclinorium carbonates, to the North Petersburg thrust zone: STOP-1 Return to cars and drive south on Route 22.
00.6	Intersection of Routes 346 and 22. Keep south on Route 22.
01.3-01.4	Large exposures of recumbently folded Synclinorium carbonates on west side of highway.
02.1	Cross trace of North Petersburg thrust fault, and proceed south on Taconic Sequence formations near base of N.P. thrust sheet.
02.2	Barn on east side of highway, house on west. We are at north edge of younger (stippled) formations at core of Church Hollow anticline (Figure 3). Bold cliffs on Taconic Mountains to east are Rensselaer Graywacke near base of Rensselaer Plateau thrust sheet.
02.8	Bomoseen Graywacke on west side of highway.
03.0-03.5	Massive exposures of Mettawee slate (subfacies -b) on west side of highway. These slates are at the core of the North Petersburg nappe.
03.5	Junction of Prosser Hollow Road and Route 22; turn east on Prosser Hollow Road.
03.7	Cross Little Hoosic River
04.4	White house on south side of highway, barn on north side. Unload for STOP-2. Walk up (west) across field to spur for exposure of Rensselaer Plateau thrust fault. Return via Prosser Hollow Road and Route 22 to North Petersburg

Mileage	
08.2	Intersection of Routes 22 and 346 at North Petersburg. Proceed east on 346.
08.5	Cross Little Hoosic River.
08.8	Turn north off 346 at intersection. Grassy low hills ahead underlain by Synclinorium carbonates; approximate trace of North Petersburg thrust fault marked by lower edge of woods.
09.0	Cross B&M Railway and Hoosic River, bear left at intersection.
09.2	Cross 10-ton-limit bridge over B&M Railway. Immediately north of bridge are outcrops of Synclinorium carbonates.
09.4	Walloomsac slate on right.
10.1	Turn right at road intersection and proceed around south end of Indian Hill (long finger of North Petersburg thrust sheet, Figure 3) on County Road 20. View to south (right) into Little Hoosic Valley with Rensselaer Plateau on west side of valley and Taconic Mountains on east.
10.7	Slates of the Owl Kill Member of the Poultney Formation on west side of road.
11.5-11.6	Cross sheared and contorted sliver of Synclinorium carbonates which marks the hanging wall of Breese Hollow reverse fault. Walloomsac slate on foot wall to west.
12.1	Turn right (east) off County Road 20, and proceed to Cipperly farm.
12.4	Unload at farmyard for STOP-3: Good exposure of metamorphosed Rensselaer Graywacke in R.P. thrust sheet, and of folded R.P. thrust plane.
12.7	Intersection of County Road 20 and Cipperly farm road. Proceed north on County Road 20. North nose of Mount Anthony visible to northeast. Rensselaer Plateau thrust fault is at base of upper steep slope. Green Mountains visible in background to north.
13.0	Walloomsac slate on west side of road.
13.4	Walloomsac slate on west side of road.
14.0-14.1	Walloomsac slate and Whipstock Breccia on west side of road.
14.5	Red barn on right (north) side of road, Synclinorium carbonates in field on south side.
14.6	Intersection of County Road 20 and NY-7. Turn east on Route 7.
15.2	New York-Vermont line. Start Vt.-9.

Mileage

16.1	Turn left (north) off Vt.-9 on Houran Road.
16.6	Unload at road bend for STOP-4; Whipstock Breccia on crest of Whipstock Hill.
17.1	Intersection of Houran Road and Vt.-9. Turn right on 9.
18.0	New York-Vermont line.
19.4	Turn right (north) off N.Y.-7 on East Hoosick Road (County Road 51). Our route now takes us into the western part of the Hoosick Falls re-entrant.
20.0	Walloomsac slate on left side of road.
20.5	Y intersection, bear left.
21.2	Road from north intersects County Road 51. Keep straight.
22.5	Walloomsac slate on right side of road.
22.7	Walloomsac slate on right side of road.
22.8	Keep straight at intersection on County Road 124. County Road 51 bears left.
23.1	View of west edge of Hoosic Falls re-entrant. The base of the hills to the west-across the valley of the Hoosick River-marks the approximate trace of the North Petersburg thrust fault.
24.3	Intersection of County Road 124 and Rt. 22. Turn right (north) on 22.
24.8	Cross bridge over Walloomsac River.
25.0	Intersection of Rts. 22 and 67. Turn left on 22.
25.05	Turn right off Rt. 22 on White Creek Road.
25.6	Unload for STOP-5 at private parking area near Little White Creek. USE NO PICKS AT THIS STOP. BE CAREFUL NOT TO DAMAGE DECKS OR WALKWAYS. FOLLOW THE LEADER. We will see a large submarine slide block of Synclinorium carbonates surrounded by Whipstock Breccia and Walloomsac slate.
25.65	Continue north on White Creek road. Cross bridge over Little White Creek. Whipstock Breccia in stream bed upstream from bridge (right side of road). Some clasts (not the cobbles in old concrete dam) in breccia are 6 to 8 inches in diameter.
26.1	Y intersection. Bear left on dirt road, then straight ahead at intersection 150 feet north.
27.4	Turn left (west) at intersection on County Road 63. Recumbently folded formations at base of North Petersburg nappe exposed on steep wooded hill to north.

Mileage

27.9 County Road 63 intersects road leading south to Eagle Bridge. Keep straight on 63. Grassy hill in foreground to south is underlain by fossiliferous West Castleton, at the base of the North Petersburg nappe. Low grassy land south of hill underlain by Austin Glen Graywacke Member of the Walloomsac. In middle distance to south is the North Hoosick klippe with trace of North Petersburg thrust fault at lower edge of woods, Austin Glen beneath the thrust, and allochthonous Lower Cambrian formations above.

28.2 Intersection of County Road 63 and Delevan Road. Grassy hill on right (north) capped by fossiliferous West Castleton limestone, and dolostone.

29.0 Intersection of Lincoln Hill road with County Road 63 at Post Corners.

29.3 Recumbent anticline (ra-1, Figure 3) in low ground to right (north), nested below other recumbent folds which are well exposed on slopes of hills in background.

29.5 Hatch Hill black slate with interbedded calcareous quartzites on right side of road.

30.2 Intersection of County Road 63 and Rt. 22. Turn right (north) on 22.

30.5 Slate of the Owl Kill Member of Poultney Formation exposed on east side of highway.

30.6 Unload for STOP-6. BEWARE OF TRAFFIC. We will see here recumbently folded Austin Glen Graywacke Member of the Normanskill, near the base of the North Petersburg thrust sheet. Gross structures best seen from west side of highway.

END FIELD TRIP

Burlington is about 110 miles to the north. Best route is 22 to Middle Granville, 22A from M.G. to Vergennes, and 7 from V. to Burlington.

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Trip B-3

EXCURSIONS AT THE NORTH END OF THE TACONIC ALLOCHTHON AND THE MIDDLEBURY SYNCLINORIUM, WEST-CENTRAL VERMONT, WITH EMPHASIS ON THE STRUCTURE OF THE SUDBURY NAPPE AND ASSOCIATED PARAUTOCHTHONOUS ELEMENTS

by

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University Park, Pa., 16802

SUMMARY

This excursion is designed to provide insight into the Paleozoic displacement and strain patterns at the juncture of the Taconic Allochthon and Middlebury Synclinorium. The nature of structural contacts will be examined in detail, and sequences of fault, fold, foliation and lineation evolution will be examined. Field trip stops will include locations within the allochthon and the synclinorium in order to place their mutual boundary relationships into proper perspective.

Assembly point: front of the office of the Vermont Structural Slate Company, near junction Routes 4 and 22A south of Poultney River in Fair Haven, Vermont. 8:30 A.M. sharp, Saturday, October 14.

Quadrangle maps: Sudbury(1946), Bomoseen(1944), Thorn Hill(1946), and Middlebury(1963) 7 1/2' topographic sheets. Most stops will be in the Sudbury quadrangle.

The two standard published references for the area of the excursions are:

Cady, W. M., 1945, Stratigraphy and structure of west-central Vermont: Geol. Soc. America Bull., v. 46, p. 515-558.

Zen, E-an, 1961, Stratigraphy and structure at the north end of the Taconic Range in west-central Vermont: Geol. Soc. America Bull., v. 72, p. 293-338.

Excursion participants may wish to review these references in advance of the field conference. Also to be recommended is the more recent review of Taconic geology by Zen (1967).

Begin Excursion: Enter Fair Haven Village and turn west on Route 4. Stop, approximately 2.5 miles at William Miller Chapel south of Route 4.

Locality 1: Taconic thrust; Structural Window at William Miller Chapel - An area of intensely-deformed Ordovician Beldens

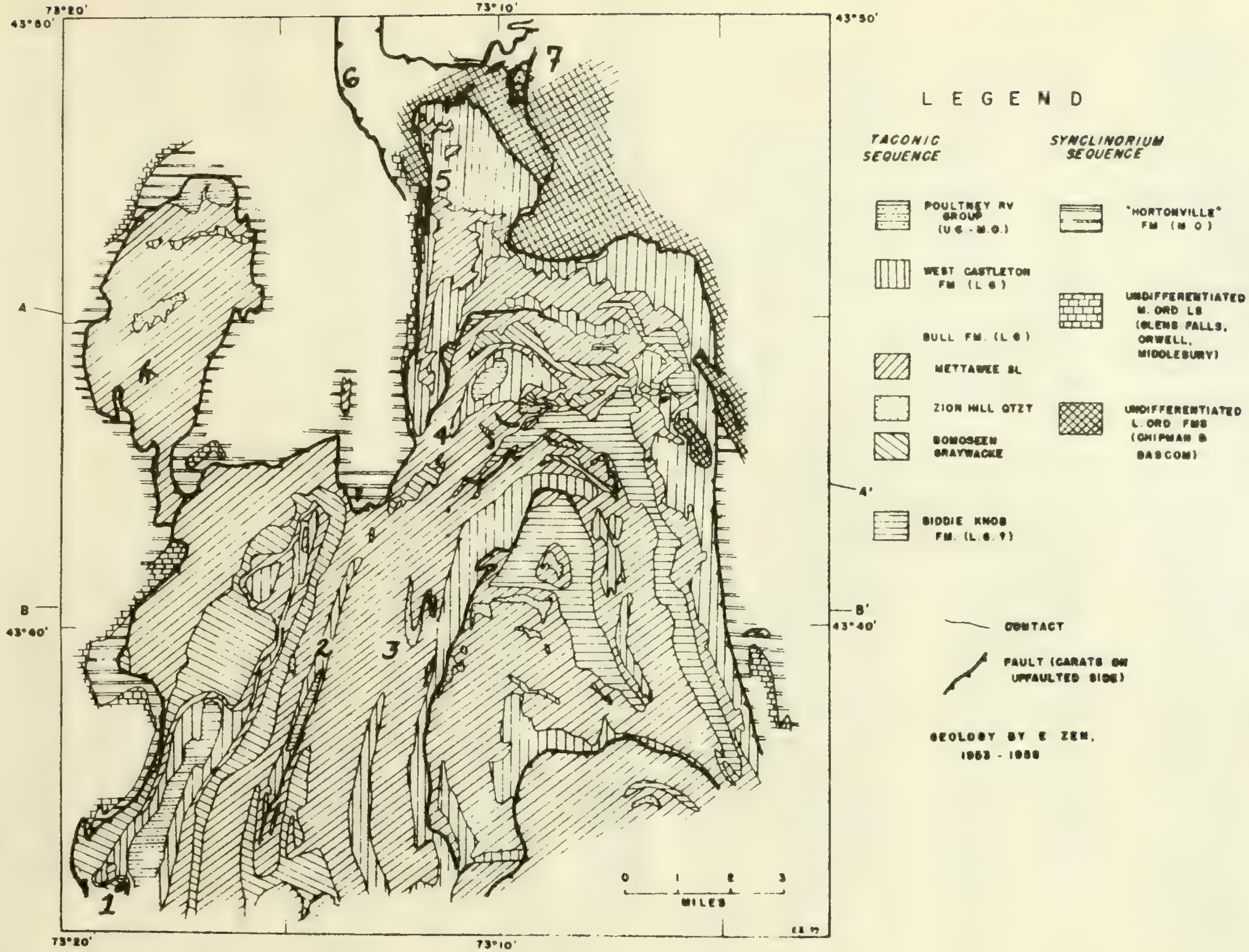


Figure 1. Geologic Sketch map, after Zen (1959), showing Excursion Localities.

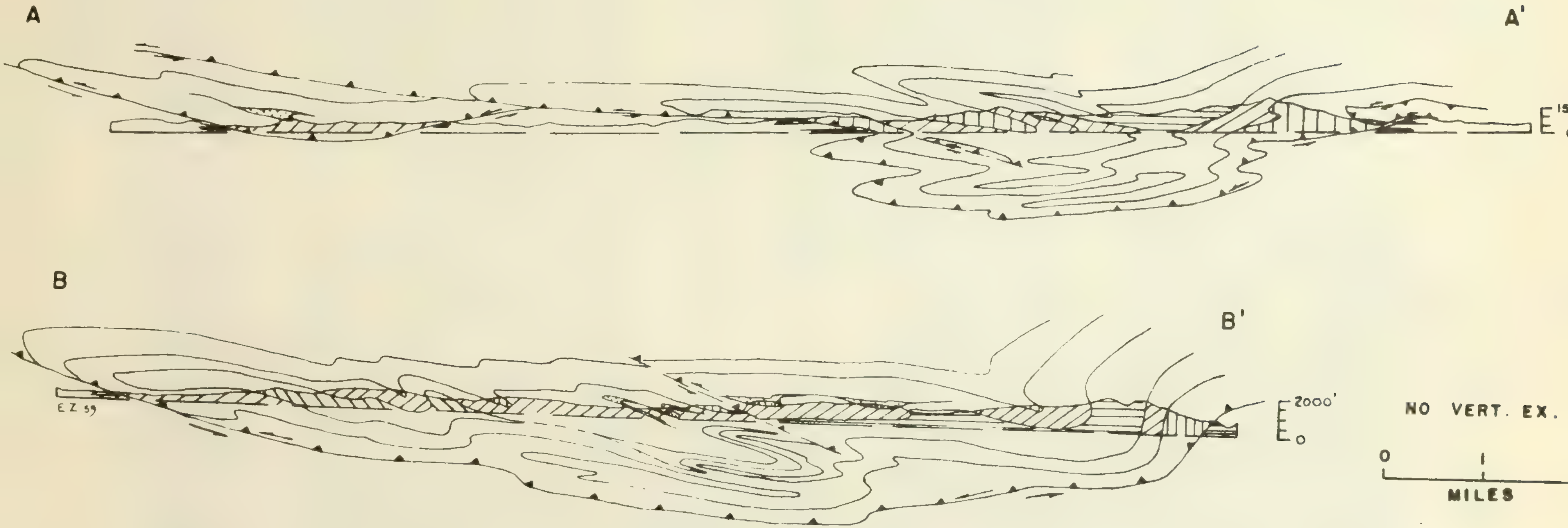


Figure 2. Generalized cross sections across the Taconic Range, after Zen (1959).

limestone and dolostone (Chipman formation, Cady and Zen, 1960) overlain in nearly horizontal contact by the early Cambrian Bomoseen graywacke member of the Bull formation (Figure 1; cf. Zen, 1961, Figure 4, p. 1.3; Plate 1). This is a structural window (Figures 1,3) in which the autochthonous carbonate is exposed through an overthrust fault sequence. The superjacent allochthonous sequence within the general vicinity is normal, showing that at least here no recumbent fold (Figure 2, B-B') exists in the Taconic sequence (Zen, 1961, p. 319; Plate 1). The structural contacts are locally exposed and can be examined in detail; there is no outcrop to the South.

The Bomoseen graywacke (Zen, 1961, p. 301-302) is typically a hard, olive-grey, coarsely-cleaved rock, weathering to white or to pale brick red. Quartz and feldspar grains, typically 1 mm, are common; white mica occurs in alignment with cleavage surfaces. A common assemblage is muscovite-chlorite-albite-microcline-stilpnomelane-quartz. Zen has presented evidence suggesting that the Bomoseen graywacke is a lithofacies that becomes progressively older and thicker to the west. It is the oldest unit in the Taconic sequence exposed west of Glen Lake.

The Beldens member of the Chipman formation is typically a white marble limestone with local interbeds of orange- to buff-weathered dolostone.

Return to Fair Haven and take the Scotch Hill road north to West Castleton.

Interlude: The first quarrying of slate in Rutland County was done by Col. Alanson Allen of Fair Haven in 1839 on Scotch Hill, about a mile north of Fair Haven village. The first quarry was worked for 8 years, using the products for hearths, headstones for cemeteries, school slates and flagging for walks, before any roofing slate was manufactured. It was one year more, in 1848, before the first roof was covered with Vermont slate. This was done by Col. Allen under the following conditions. He was to wait for one year for his pay, and if, in the meantime the roof should break down from the weight of the slate, he was to receive no pay, but should pay all damages. The farmer was disappointed and the roof is good today (Smith and Rann's, History of Rutland County, 1866). The barn still stands on the farm of Stanley Kruml, about a mile south of Fair Haven on Rt. 22-A; the roof is in excellent condition.

Locality 2: Scotch Hill Syncline, West Castleton - This is private property. We are permitted to be here by the courtesy of the owners. Sampling of rock specimens is not permitted; LEAVE ALL HAMMERS IN THE VEHICLES ! Please cooperate -- thank you.

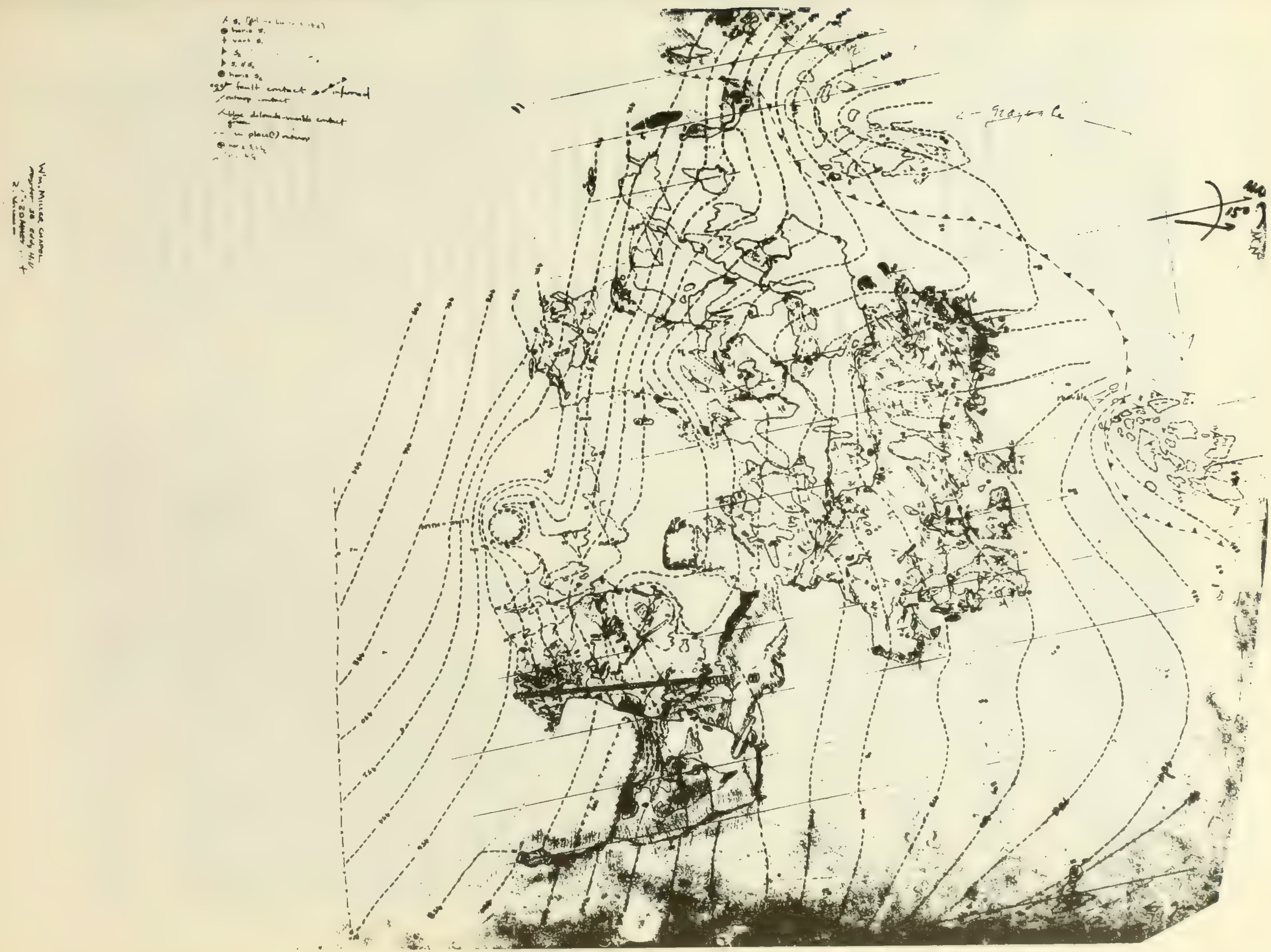


Figure 3. Plane-table outcrop map of structural window, William Miller Chapel.

Exposed cross-section of the Scotch Hill syncline: the east limb is nearly vertical at this locality (Figure 4). The west limb dips east at a shallow angle. This is the West Castleton formation (Zen, 1961, p. 304), within which early Cambrian fossils have been found by several workers, in ledges near this locality.

The West Castleton formation ranges from a dark-grey, hard, poorly-cleaved sandy or cherty slate that weathers white or pale-red to a jet-black, fissile, graphitic and pyritic slate that contains many paper-thin white sandy laminae and commonly also black cherty nodules, and when weathered displays much alum bloom. Locally interbedded in the fine black slate are beds of buff-to yellow-weathering black dolostone or dolomitic quartzite, a few inches thick, some of which, however, become massive, siliceous, and heavily bedded in the harder black slate. The varieties of black slate do not form mappable units but grade into each other along strike (Zen, 1961, p. 304-305). The rock becomes a phyllite to the east.

Immediately to both the east and west of this fold are quarry belts of green and purple slates (Mettawee member, Bull formation). The simplicity of the Scotch Hill structure and the relation to the underlying Bull formation is consistent with the relative ages of the two units as proposed by Zen (1961, p. 317). The flexural flow fold displays a host of well developed minor structural elements, e.g. cleavage in several rock types, "refraction" of cleavage, slip and flow phenomena in the dolostones; calcite-filled fractures; and pressure-shadows associated with porphyroblasts. Cleavage is axially-planar in thicker slate beds.

The dating of the cleavage is an important but unclosed question; Zen (1967, Appendix 7, p. 101) has suggested a post-Ordovician age of the regional metamorphism based both on regional relations and on radiometric dating. Zen noted that the regional metamorphic grade increases steadily eastward, from the non-metamorphosed rocks of the Champlain Valley, to almandine-kyanite grades in eastern Vermont. No multiple-metamorphic effects were observed in the Paleozoic sequence; rocks as young as Devonian are involved and the metamorphic episode was presumed to postdate this. Radiometric dates in the range 350-390 m.g., from the southern Taconic region and from north-central Vermont (Camels Hump area) are cited by Zen in support of this view. Yet Harper (1968) has cited radiometric dates from slate belt minerals in the range 445-460 m.g., which dates would be compatible with Ordovician (Taconic) deformation. This problem remains unresolved, and demands attention; a correct interpretation of regional structural relationships within the "Taconic" and "synclinorium" areas hinge upon its resolution.

Drive eastward along the Cedar Mountain Road to the (dead) end. Quarry on the left at the end.

Locality 3: Cedar Mountain Syncline; slate quarry - This, the first major structure west of Lake Bomoseen, is spectacularly exposed in the abandoned quarry cut into Cedar Mountain (Figure 4). Fold is overturned to the west, nearly recumbent, and plunges south; axial plane cleavage dips gently eastward. The deformational mechanism associated with the exposed portion of the fold, appears to be (chiefly) passive flow; on a larger scale, flexural mechanisms may have played a role. Exposed rocks are in the upper beds of the Bull formation, Mettawee slate facies (Zen, 1961, p. 300-301), chiefly a soft purple and green slate with local thin beds of limestone.

Typical mineral assemblages are muscovite-chlorite-albite-quartz and muscovite-chlorite-hematite-quartz (Zen, 1961).

The anticline separating location 2 and 3 is not well exposed here, being for the most part masked in the pervasive cleavage of the Mettawee slates; its presence, however, is indicated by exposures of Bomoseen graywacke southwest of Lake Bomoseen (Zen, 1961, Plate 1).

The purple and green colors of these slates reflect different relative proportions of chlorite and hematite. Certain color features bear on the question of whether or not chemical equilibrium was obtained during conditions of metamorphism (Zen, 1960, p. 167): thus it may be observed that hematite-bearing purple slates are never found in contact with graphite-bearing black slates without an intervening layer of green slate and that pyrite-spots and limestone layers in purple slates are always surrounded by a rim of green slate. These layers are reaction rims; this evidence, together with textural data and the observation that mineral assemblages obey the phase rule, suggest the rocks have, in the main, achieved chemical equilibrium during metamorphism (Zen, 1960).

Within the cleavage plane a subtle lineation can sometimes be observed; this is termed "grain" by quarrymen. Grain has been observed to be roughly perpendicular to fold hinge lines; it appears to be formed from a preferred orientation of elongate mineral grains, although definitive evidence on microstructure of the feature has not been reported. As a working hypothesis, the writer has believed "grain" to be the direction of greatest finite extension within the plane of flattening; this view remains hypothetical. Nonetheless the feature may ultimately prove to be an important one in structural analysis. In support of this view, Wright (1970, p. 55) observed ellipsoidal green reduction spots in purple slates, and concluded that the longest dimension was parallel to the grain direction. Reduction spots appear to be reliable as strain indicators, and the implication is that "grain" may also be a useful strain indicator. The reduction spots have long and medium dimensions within the plane of flow cleavage; flow cleavage is thus presumed to have developed perpendicular to the axis of greatest finite

Figure 4. Location Map, Localities 2 and 3; Bomoseen $7\frac{1}{2}$' Quadrangle.

compressional strain. It should be noted that this direction is in general not equivalent to the axis of maximum compressional stress. Maximum values of analyses of flattening in the Vermont slate belt, based on reduction spots, is approximately 80% (Wright, 1970, p. 64).

Larrabee (1939-40) and Dale (1899) have discussed the principles of structural geology in relation to flagstone and slate quarrying. Within the slate belt of Vermont and adjacent portions of New York State, slate and flagstone are quarried from three rock units; the Mettawee Slate, the Poultney Slate, and the Indian River formation. Mettawee yields purple, grey, and rarely grey slates; Poultney slates are grey-green, and the Indian River red and blue-green slates. The latter is quarried only in New York, near Granville.

Interlude: To the south of the quarry is Neshobe Island, formed principally of slates of the West Castleton formation, but locally containing the Beebe limestone member of the West Castleton formation, a massive, lenticular black limestone. A thin band of Mettawee slate occurs at the eastern extremity of the island. Of greater interest perhaps is the observation that author Alexander Woolcott owned the island in former years, and that such diverse fauna as Marx Brothers were known to have prowled through its lush undergrowth.

Due east, on the crest of the Taconic Range, is the 1976-foot peak of Grandpa Knob. Here, in 1941-45, Palmer Putnam's 1500-kilowatt wind turbine made electrical research history. A 150-foot windmill with stainless steel blades generated power that was fed into utility lines of the Central Vermont Public Service Corporation, Rutland. A technical success, financial obstacles hindered further development when, in 1945, one of the eight-ton blades had been ripped from its shaft and tossed 750 feet down the mountain.

Return to West Castleton; turn north on Moscow-Black Pond-Hortonville Road to its intersection with the Seth Warner Memorial Highway (Route 30). A few miles to the southeast of the road intersection is the site of the Battle of Hubbardton, the only contest fought on Vermont soil during the American Revolution, on July 7, 1777. A museum at the site is contained within an 18th century-style building; it features an animated electrical relief map and diorama depicting the important stages of the battle. (The Battle of Bennington took place on New York soil). Turn north along Route 30, for about 0.5 miles. Stop south of Eagle Rock camp (parking may be difficult).

Locality 4: Structure of the Giddings Brook "Slice": Problem of the Giddings Brook-Ganson Hill Fold Complex - In the

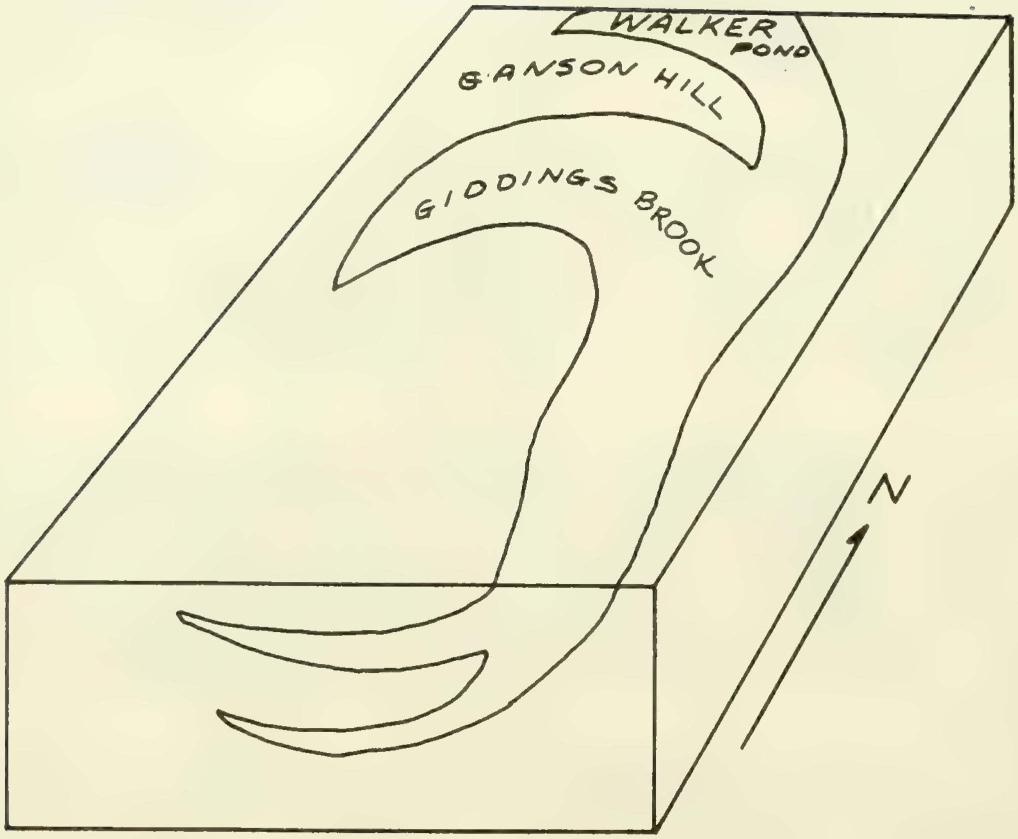


Figure 5. Schematic diagram showing the geometry of the Giddings Brook bottoming fold as proposed by Zen (1959, Pl. A-4; 1961, Fig. 4). Surface shown is the top of the Biddle Knob formation. Topographic effects are ignored. The structure is shown to be recumbent with a shallow South-plunging hinge line.

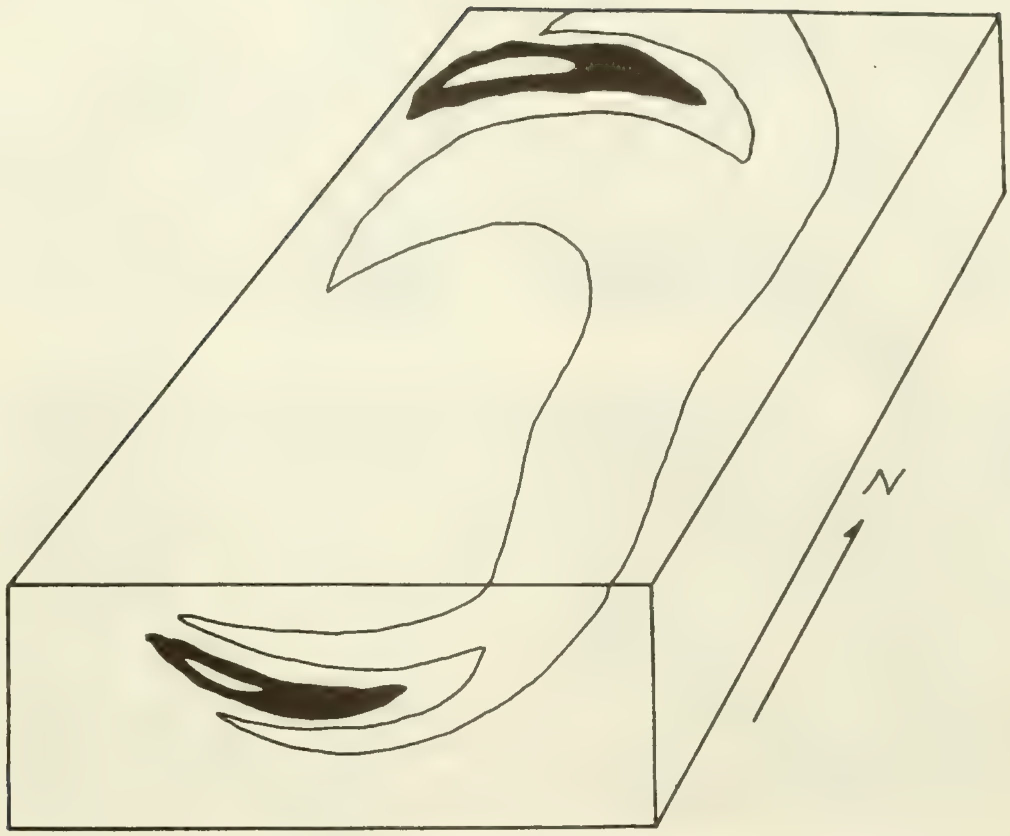


Figure 6. Schematic diagram showing the geometry of the hypothetical Giddings Brook bottoming fold with the outcrop pattern of West Castleton formation and younger rocks included in plan view, and its inferred profile.

east-central part of the geologic map (Figure 1) there is a large, boomerang-shaped tract of Biddie Knob formation, and a half-moon shaped area of the West Castleton formation and the Poultney River group immediately to the northwest (Zen, 1959, p. 3); these represent the Giddings Brook-Ganson Hill fold system of Zen (1961, p. 316), in which the Biddie Knob formation is assumed to form the core of recumbent "bottoming fold"* along the Giddings Brook valley, and the West Castleton and younger rocks are assumed to be contained within a recumbent syncline, the Ganson Hill syncline (also a "topping" fold). Locality 4 in Figure 1 occupies the western tip of the Ganson Hill syncline. The schematic diagram shown in Figure 5 (cf., Figure 2, A-A') provides a clear picture of the structural geometry as envisaged by Zen (1959, Plate A-4; 1961, Figure 4).

Such a structure seems plausible if only the Biddie Knob formation is considered; but in my view it seems implausible when, in addition, West-Castleton and younger rocks in the vicinity of Ganson Hill are considered. The Ganson Hill syncline exhibits closure at its western end near the Seth Warner Memorial Highway, as previously shown by Zen (1961, Plate 1; p. 316) and other workers. This would require a schematic illustration something akin to Figure 6, which in turn suggests the "cylindrical" structure portrayed in Figure 7. Such a structure would be explicable only by large-scale boudinage; this latter hypothesis does not appear to fit the field evidence, which evidence suggests that the Ganson Hill structure is a comparatively shallow overturned syncline with a flat northeast-trending hinge line, comparable to and possibly an extension of the Scotch Hill syncline (Locality 2).

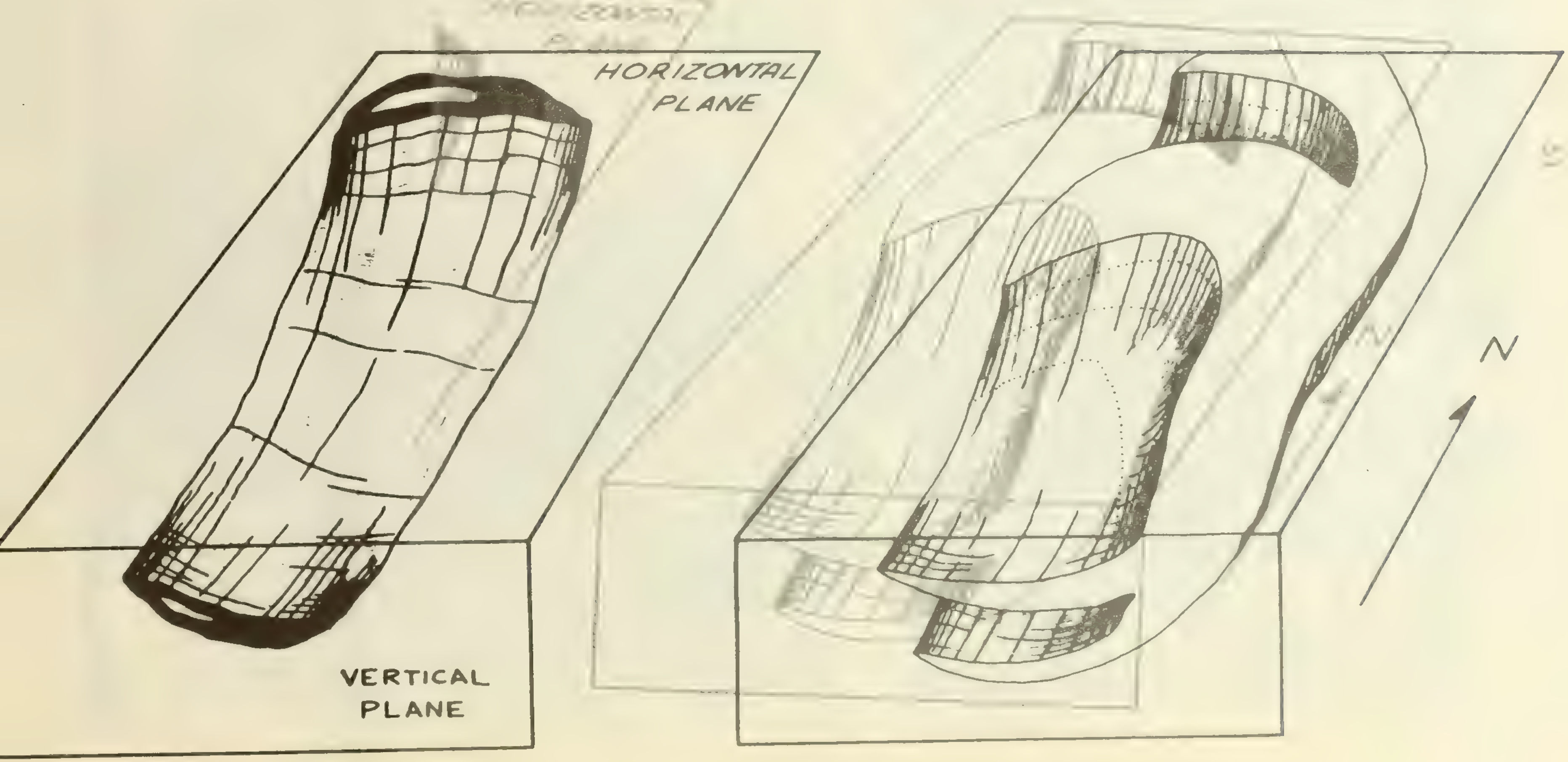
Continue north on Route 30 to Sudbury.

Interlude: Church at Sudbury Village. Built in 1807 and later granted joint use by town and the congregational services, the exterior shows in its design the lingering tradition of "gothic" detail of Old England's churches. The former galleries have now been replaced by a floor, the Town Hall being on the ground floor and the religion services upstairs.

Take the road southeast from Sudbury Church for 0.6 mile to sharp turn with road cut. Go through fields westward to exposures of white marble, crossing a "Taconic thrust"; follow the contact southward.

*Zen (1961, p. 313-314) introduced two new terms in order to describe the structural complexity of this area: "By topping fold is meant a fold whose core contains the relatively youngest beds. By bottoming fold is meant one whose core contains the relatively oldest beds. For rocks that have only been simply folded, these terms are equivalent to synclines and anticlines, respectively; however, for rocks which have been complexly deformed, these terms are not necessarily synonyms...topping- and bottoming folds are terms with stratigraphic connotations."

Figure 7. Schematic diagram showing individual structures according to the recumbent Giddings Brook-Ganson Hill fold complex of Zen. On the right, the structure inferred for the Biddle Knob formation. On the left, the "cylindrical" structure "required" for West Castleton and younger rocks.



Locality 5: Taconic Margin east of Hyde Manor and Sudbury; Historical Ground for Taconic Controversy - Here Taconic sequence rocks of the "Signal Hill Slice"(viz., all Taconic sequence units north of the Keeler Pond fault of Zen (1961, Plate 1; cf. Voight, 1965) overlie parautochthonous rocks of the Sudbury nappe in imbricate thrust contact (Figures 8, 9, 10; cf. Kay, 1959). The field relationships here suggest that Arthur Keith's original evidence for the Taconic klippe, which had long since fallen into disrepute, was basically correct. Keith (1913) maintained that an overthrust was indicated because the limestones and slates were unconformable with the general contact: "Inasmuch as the divisions of the Ordovician Stockbridge limestone in this area dip under the slates known to be Cambrian and these in turn dip and pitch away from the limestone, and inasmuch as the limestones and slates are all unconformable with the general contact, an overthrust seems the only competent explanation." Keith was correct; the fact that Keith probably misidentified the infolded black Hortonville slate, and the possibility that local occurrences of so-called "Taconic sequence" slates may also be Hortonville, have little bearing on this general conclusion. This is a thrust contact, not simply an unconformity as had been contended by some subsequent workers.

Also of historical interest is an "outlier" described by T. N. Dale, one of the pioneers of Taconic geology. In his 1904 paper describing the geology of the northern Taconic region, Dale cited an "outlier" west of Hyde Manor as evidence that the carbonate sequences lay unconformably on top of the slate-phyllite Taconic sequence rocks. Subsequently Ruedemann (1909), in the first suggestion that a "Taconic thrust sheet" underlay the slate belts, cited Dale's "outlier" as a true "fenster" and called it "positive evidence" for the overthrust. Ruedemann claimed that the limestone was an anticline protruding from below, rather than a syncline as visualized by Dale. The Ordovician age of the "outlier" limestone was known from fossil evidence (streptelasma, crinoids) as reported by Dale. Valuable as is the use of imagination in geological investigations, geological sciences are still best advanced by careful observation and deduction; thus heeding his own words, Dale advanced in 1910 on the outcrop and, "with the aid of two men and dynamite" made six excavations; in 1911 Dale drilled a core through the center of the outcrop that penetrated the carbonate layer at a depth of 14 feet (Figures 11, 12). Ruedemann's fenster theory was thus deemed unlikely (Dale, 1912), but not invalidated completely, for an overturned anticline with an east-dipping axial plane could be compatible with drill core data. Hence Dale (1913) continued his study of the locality, increasing the number of excavations to fifteen, and drilled another core, inclined 45°, roughly parallel to the axial plane, which passed through limestone at about 32 feet (Figure 11). These papers by Dale might well be considered classic but in fact do not seem to be well known.

Cushing and Ruedemann (1914, p. 113) then admitted the likelihood of the outcrop as representing a small infolded mass, but at-

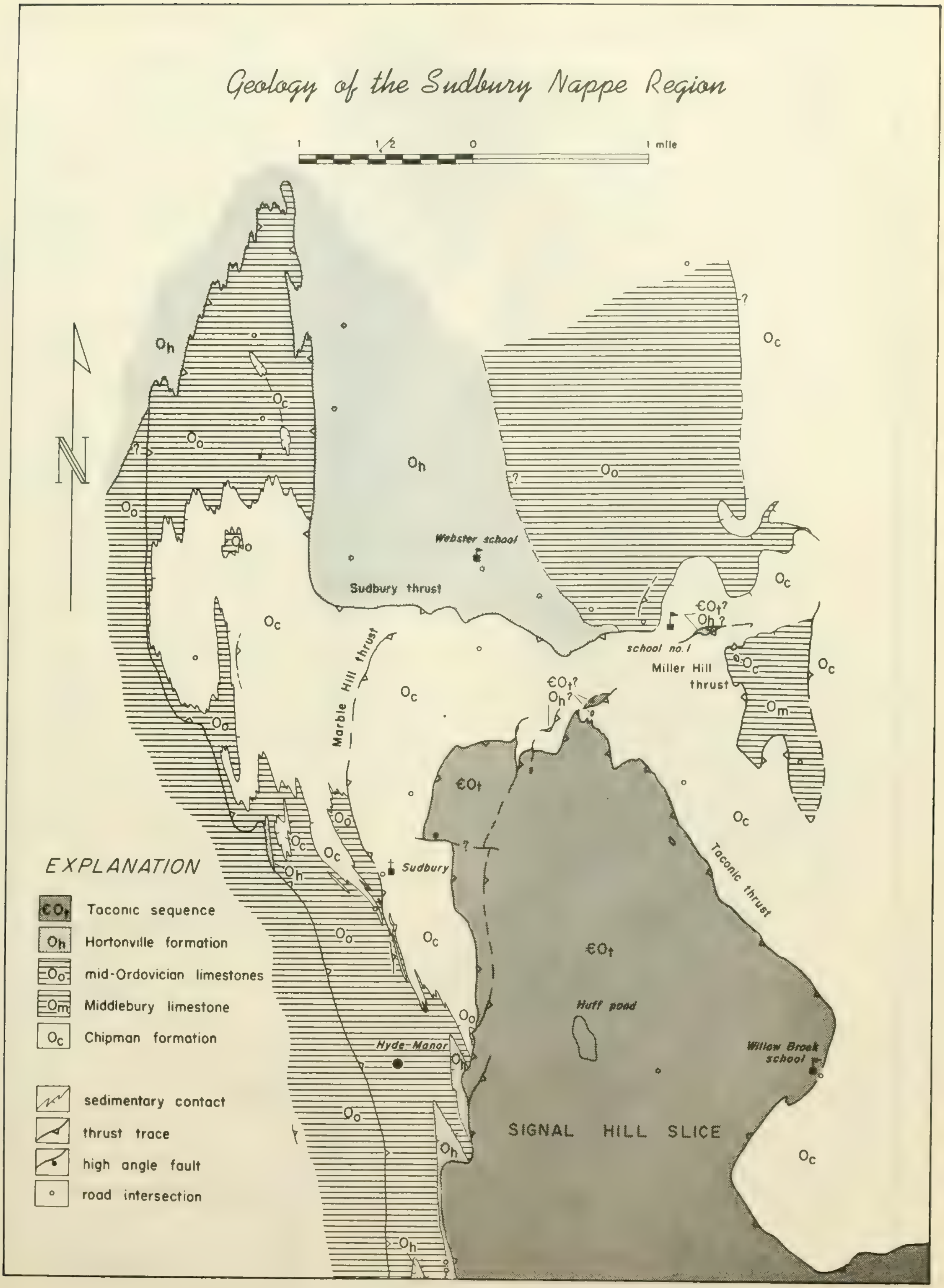


Figure 8. Geology of Sudbury nappe region.

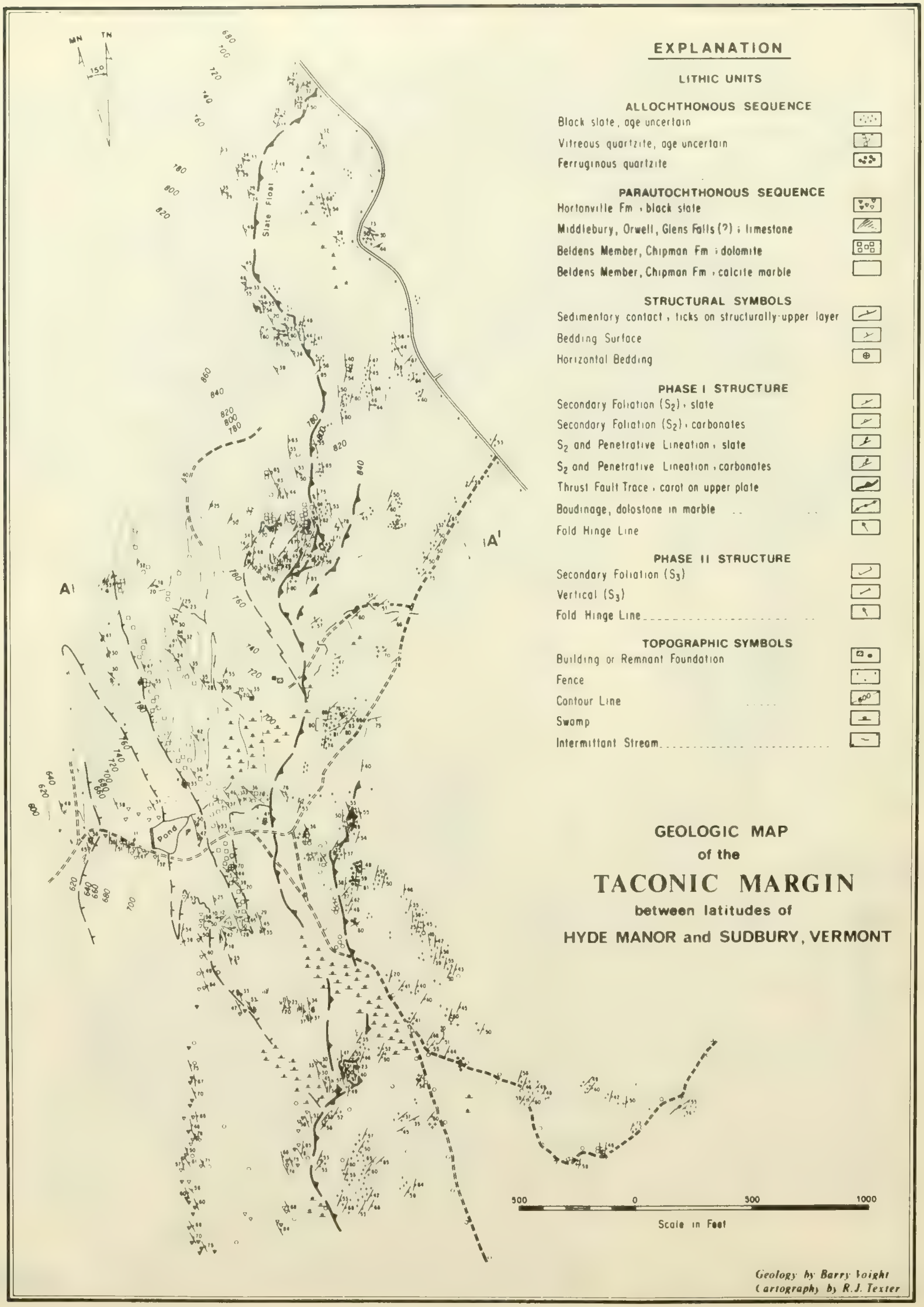


Figure 9. Geologic map of the Taconic Margin between latitudes of Hyde Manor and Sudbury, Vermont.

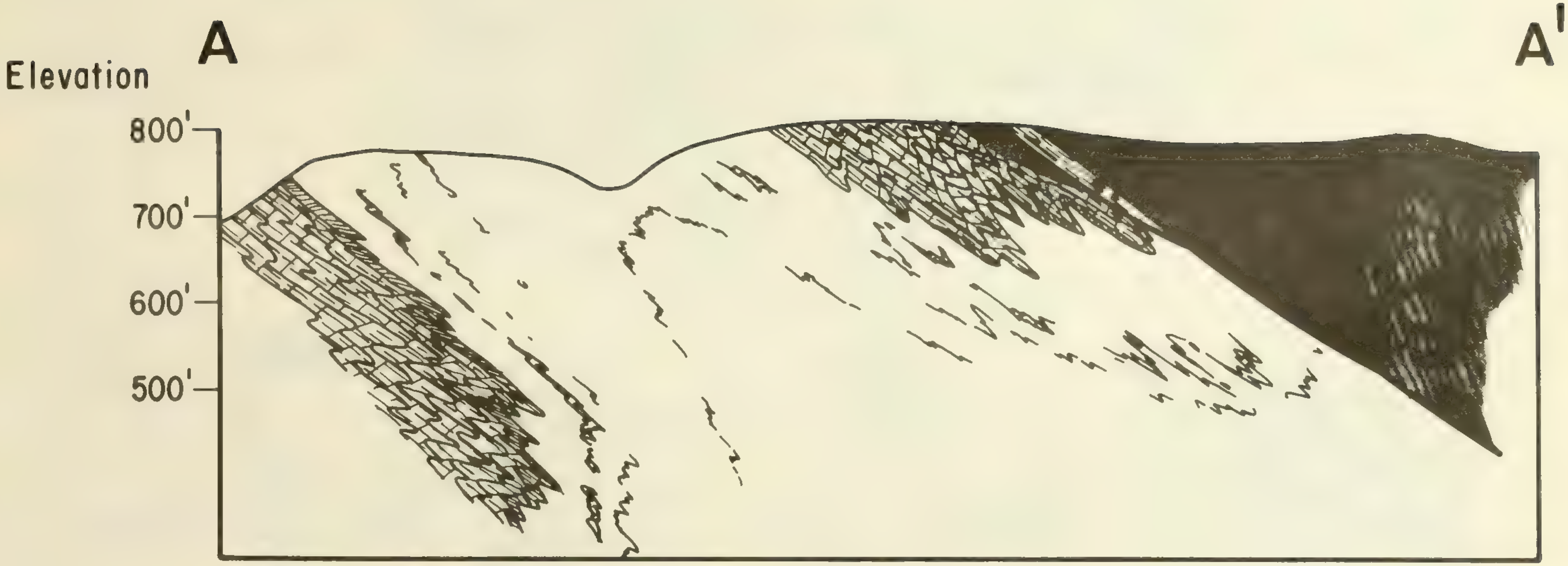


Figure 10. Cross section of Taconic margin; cf. Fig. 9

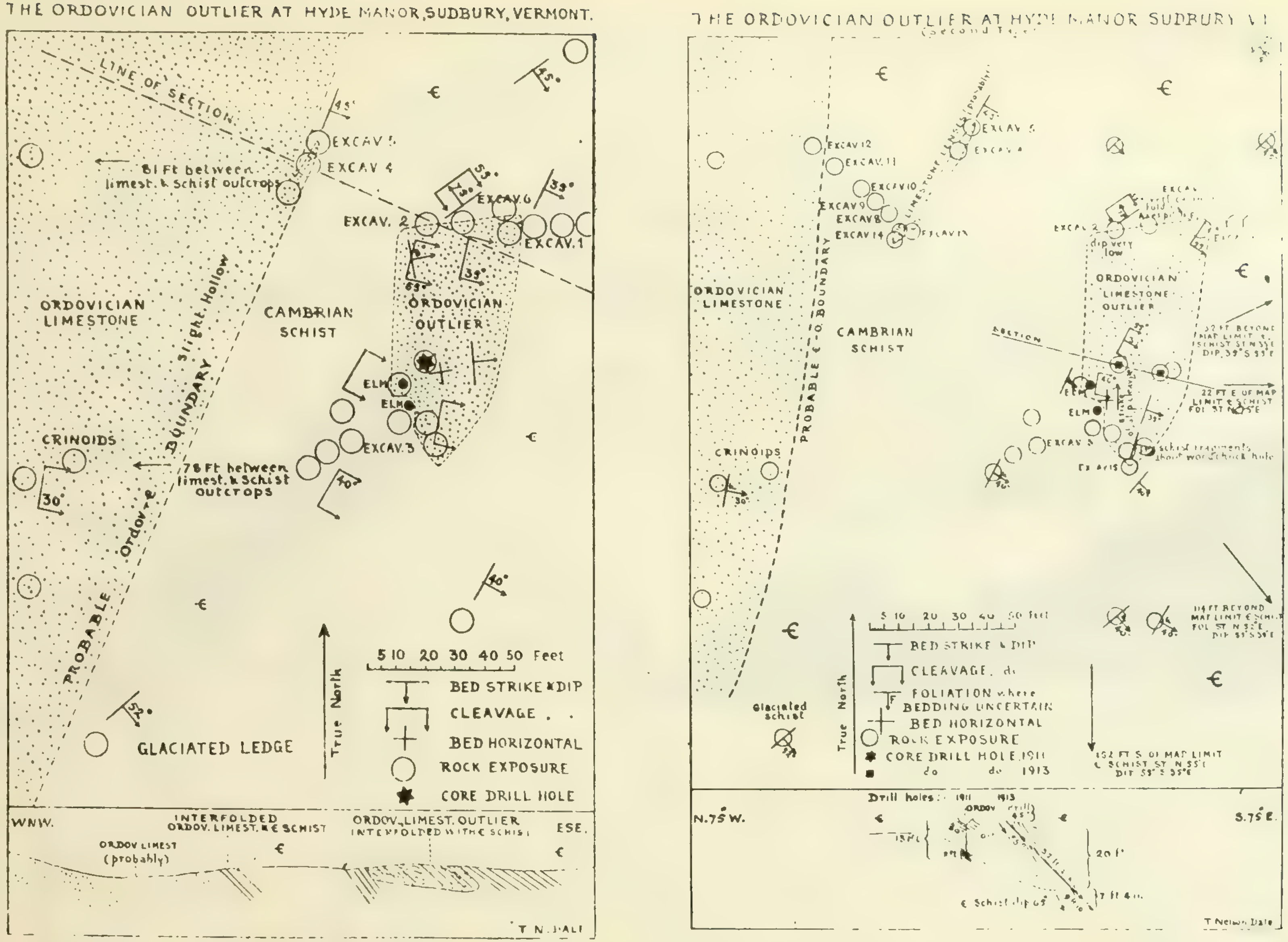


Figure 11. Ordovician "outlier" of T.N. Dale; left, 1912 map; right, 1913 map (second paper).

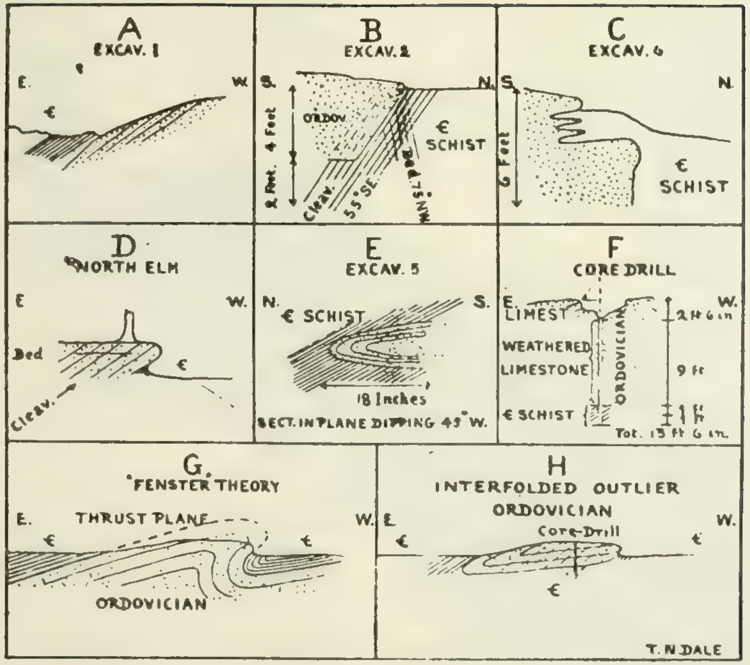


Figure 12. Structural details and alternative hypotheses; excavation numbers refer to the 1912 map of T.N. Dale (Fig. 10).

tempted to diminish its importance because of its small size. Kaiser (1945, p. 1095) commented that Dale's investigations were of historical value only, since present workers "agree that the slate is on top of the limestone here"; he discounted the drill core evidence. Keith was at the outcrop with Dale when field work for Dale's 1912 paper was in progress, but never mentioned it in his brief accounts of the geology of the region (1912, 1913, 1932, 1933) although he must have considered it in evidence. Most other workers have not cited it. Notwithstanding the above, a multiple working-hypothesis approach is effective only if a sufficient number of hypotheses are included in those to be tested. In this instance the "outlier" seems to be but one of several located near the "Taconic" margin; more recently they have been interpreted, not as "outliers" or "fensters", but as carbonate slivers caught up in imbricate thrust slices (Voight, 1965). The "outlier" of Dale is the most southerly of the tectonic "lenses" shown in Figure 9. Similar imbricate relationships have been observed at the northern tip of the "Taconic" margin (cf. Figures 8, 13, 14).

Interlude: The FIFTH, who chanced to touch the ear,
Said: "E'en the 'blindest man
Can tell what this resembles most;
Deny the fact who can,
This marvel of an Elephant
Is very like a fan !

Return to Route 30 via Sudbury; turn north on Route 30 approximately two miles to the Webster School, and turn left toward Orwell. Stop about 2.3 miles to the west, before the Lemon Fair River Bridge. Cross fence (keep all gates closed !) and enter fields to the north of the road.

Interlude: Any wandering movement would have to occur across the mountain chain...The "axles" and "rollers" would have to operate parallel to the length of the mountain wall, and the guiding tracks would have to run at right angles to it.

When we began to use the compass and plot the measurements on the map, we found that our expectations were being met only in respect to the glide tracks. All signs of a rolling motion were at right angles to the direction we had expected. In other words, axles and track ran parallel !

We were in the position of the engineer who stands on the railroad tracks, and sees a locomotive travelling towards him. As it draws closer he suddenly discovers that the wheels are placed crosswise to the track, and that the axles run parallel to the rails. Therefore it is obvious that the machine cannot roll. Yet it does ! Should he jump to one side and out of the way ? Or shall he trust his theory that motion is impossible under such conditions and stay

Figure 13. Imbrication at the northern tip of the Taconic Range (north Fiddle Hill area).

where he is, and perhaps get run over ? We jumped. But some five years later we regretted this cowardice, and returned to stand on the rails...

There were, however, still many unsolved problems. Is it strange that we were unable to comprehend the behavior of a mountain range 625 miles long in terms of lathe and locomotive ?

Locality 6. Boudinage, Lineation and "Early" Folding, Lemon Fair River Bluffs - LEAVE ALL HAMMERS IN THE VEHICLES. All relevant features can be seen best on natural surfaces; the outcrop is unique and should be preserved from death by percussion.

Bluff exposures are of interbedded white to grey marble and buff-to-brown ("chamois weathering") massive and "cleaved" dolostone, typical of Beldens member, Chipman formation, and structurally near the base of the Sudbury nappe. These rocks have been deformed into a magnificant cascade of nearly-isoclinal (early) folds with associated structural features (see map, Figure 15). A synoptic diagram of structural data for this locality (Station 7 of Voight, 1965) is given in Figure 16; early fold hinge lines trend toward the southeast at a low plunge, fold axial planes and associated secondary foliations dip eastward at a shallow angle, penetrative lineations within the marble dip eastward, neck lines of boudins plunge northward at a shallow angle.

Boudins thus have a different trend than early fold hinge lines, and are somewhat younger structures than the fold hinges; locally fold hinges are pinched-off, although maximum development occurs on fold limbs (Figure 17). Nonetheless, boudins are assigned to an early deformational event; their evolution appears to be related to the development of the secondary foliation associated with the early folds. The enormous ductility of marble inferred from the boudins is characteristic of early deformation, and development of boudinage can be logically envisaged as the result of continued compression of isoclinally folded structures. A detailed sketch of boudins at this outcrop is given in Figure 18, with numbers assigned with reference to Figure 15. The smoothly-tapered boudin geometry had earlier led one mis-informed geologist to conclude that the massive dolostone layers were extremely ductile during conditions of deformation (Voight, 1964a; 1965), a conclusion having some significance with regard to the inferred deformational environment and to the inferred mechanics of formation. However, inspection of the boudins themselves shows clearly that the mode of deformation of the dolostone layers has been by pervasive brittle fracture and fragmentation; the pseudo-"rounded" boudin geometry appears as a consequence of fragmentation and (predominantly) calcite vein filling. The characteristic buff-weathering of the dolomite tends locally to mask the degree of fragmentation and vein filling in boudin necks, although the "thread-scored beeswax" patterns on the weathered surface are, in point of fact, fracture trac-

Figure 14. Imbrication at the northern tip of the Taconic Range (Confusion Ridge area).

Figure 15. Map of a portion of Locality 6, showing location of boudins (modified from Crosby, 1963). Black and stippled beds are dolostone; white is calcite marble.

Figure 16. Synoptic diagram of structural data, Schmidt Net, for Locality 6. Fold hinge lines (small circles), pole to πS_1 (large circle), pole to πS_2 (crossed circle), neckline 13% maxima (lined pattern), bedding pole (S_1) 10% maxima (dotted pattern), secondary foliation pole (S_2) 10% maxima (crosses), penetrative lineation 27% maxima (triangles).

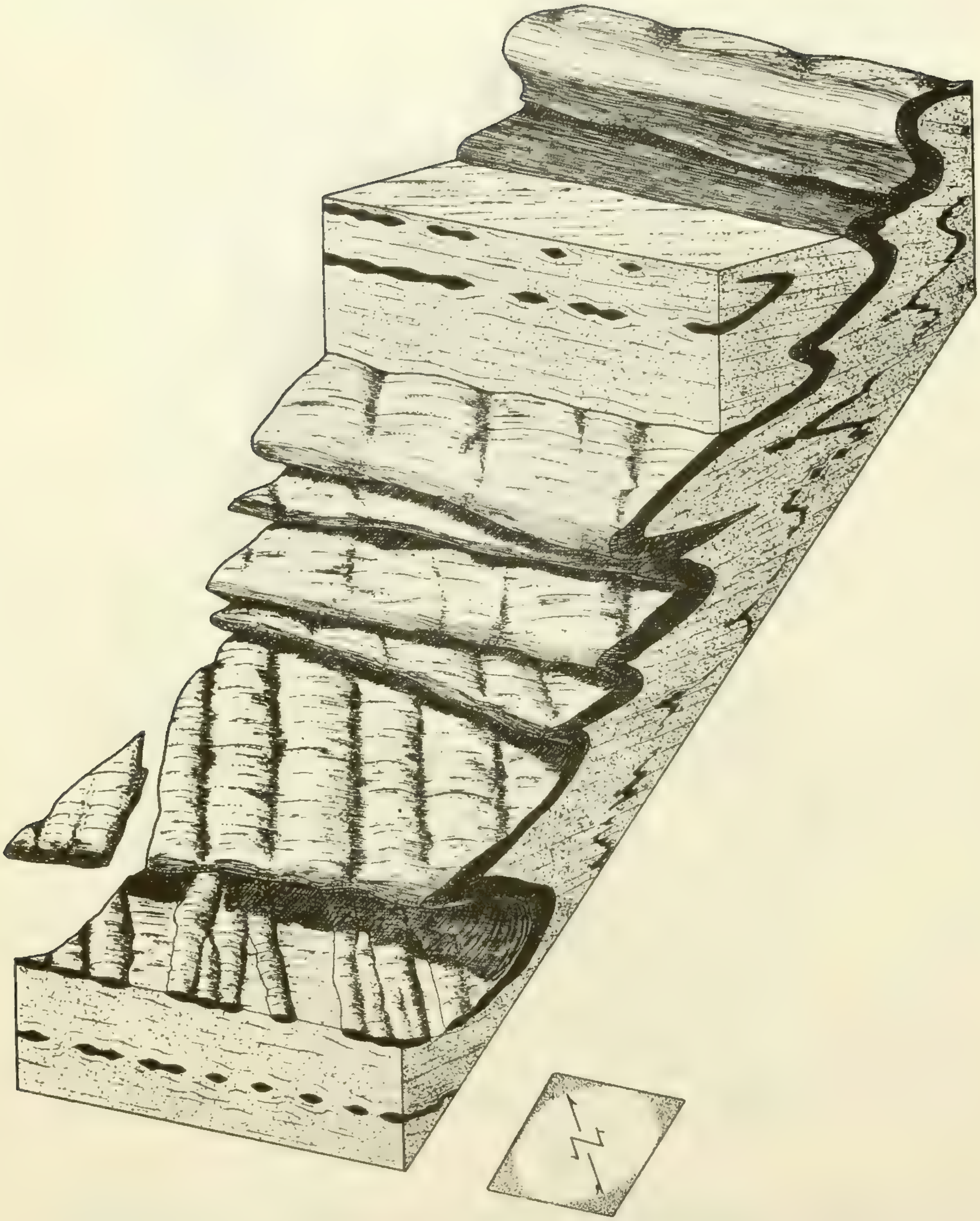


Figure 17. Schematic block diagram of isoclinal folds, boudinage, foliation, penetrative lineation for Locality 6.

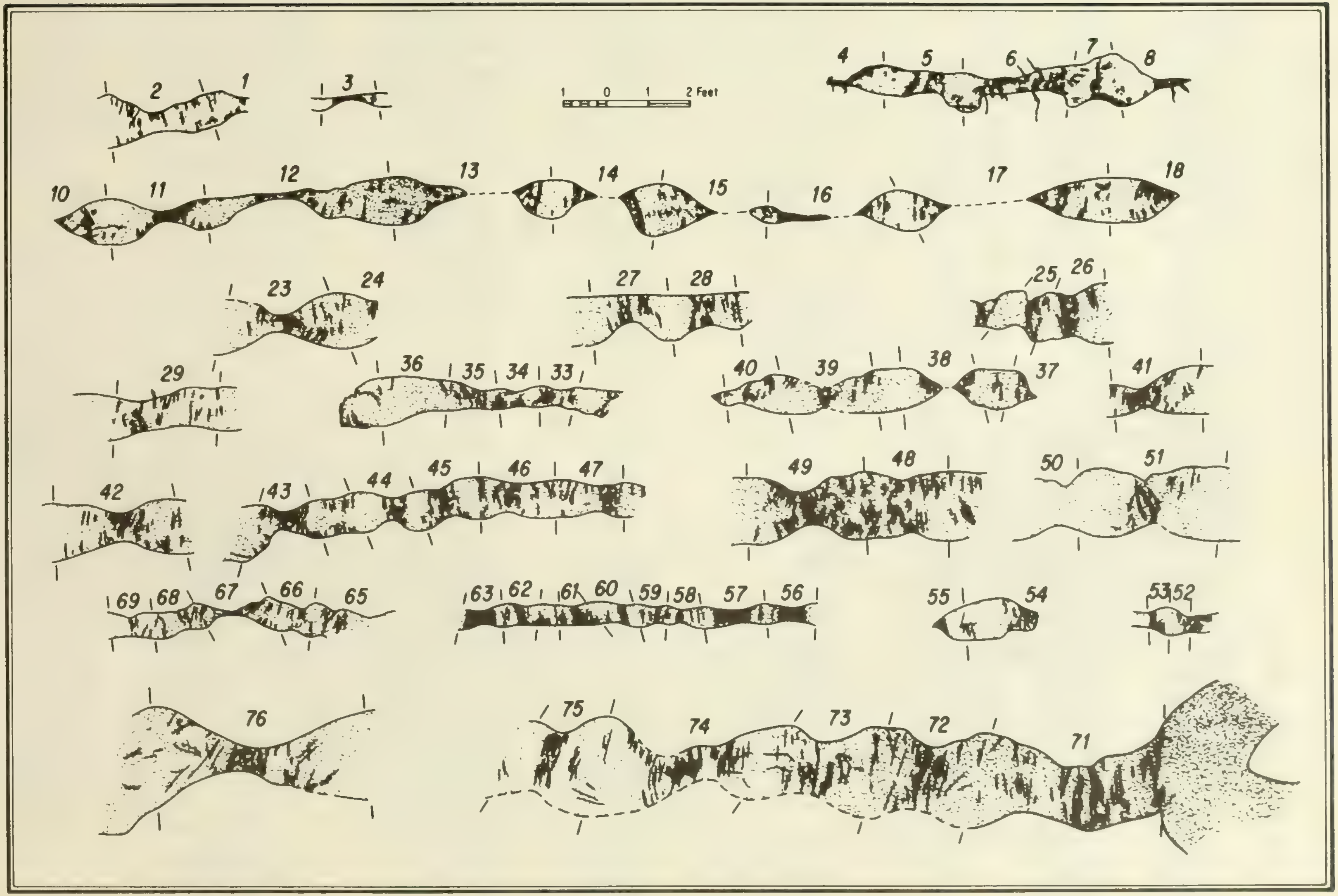


Figure 18. Transverse boudin profiles, dolostone in marble matrix; Locality 6; cf. Fig. 15. Calcite veins are shown in block.

es on that surface.

Boudins can be utilized in strain and ductility measurements according to a simple procedure outlined by Voight (1964a). In the description here, two-dimensional analysis is assumed appropriate.

The neck (constriction) separating contiguous boudins is treated, assuming zero axial strain, constant volume in deformation, and an original layer thickness (t_o) not less than maximum boudin thickness. Original neck width (W_{n_o}) is given by

(1) $W_{n_o} = A_n/t_o$, where A_n represents neck area, easily derived from

planimeter-measured tracings of down-axis boudin photographs. Average strain at the neck (ϵ_n) obtains:

(2) $\epsilon_n = (W_n - W_{n_o}) / W_{n_o}$; the existing neck width is W_n.

If boudins are isolated, their separation is indicative of matrix strain; minimum values may be computed. By restoring boudins to their positions at initial separation, average boudin-layer strain is determinable. If A_i equals individual boudin area in an array, the equation:

(3) $W_o = (\sum_{i=1}^{n} A_i) / t_o$

provides width of an assumed original layer of constant thickness. Coupling this dimension with the extended width (W) of the boudin layer gives:

(4) $\epsilon = (W - W_o) / W_o$.

Longitudinal strain values for both boudin and matrix layers provide data on rock ductility during deformation. The limiting thickness ration, t_n/t_o, where t_n is the minimum thickness of the neck, also serves as a measure of ductility for boudins which have developed as a consequence of continuous flow. An example of this kind is given in Figure 19.

Total cross-sectional area for necks 11 to 17 is about 11.5 sq. ft.; cumulative width is 24.1 ft.; the "best estimate" of initial thickness is 1.1 ft. From equation (3), cumulative initial width is 11.5/1.1 = 10.5 ft. Minimum(local) extensile strain in the marble "matrix", from equation (4), is ((24.1 - 10.5)/10.5) 100 = 131%. Restored width at incipient separation is about 19.7 ft.; hence average dolomite (boudin layer) strain is ((19.7 - 10.5)/10.5) 100 = 88%. In this example no corrections were made for vein fillings, hence the calculated values are minimum values.

Note that the spacing between necks is reasonably consistent within individual layers, but extremely variable when one layer is compared with another. The critical factors controlling spacing include both material properties and geometric properties; in a

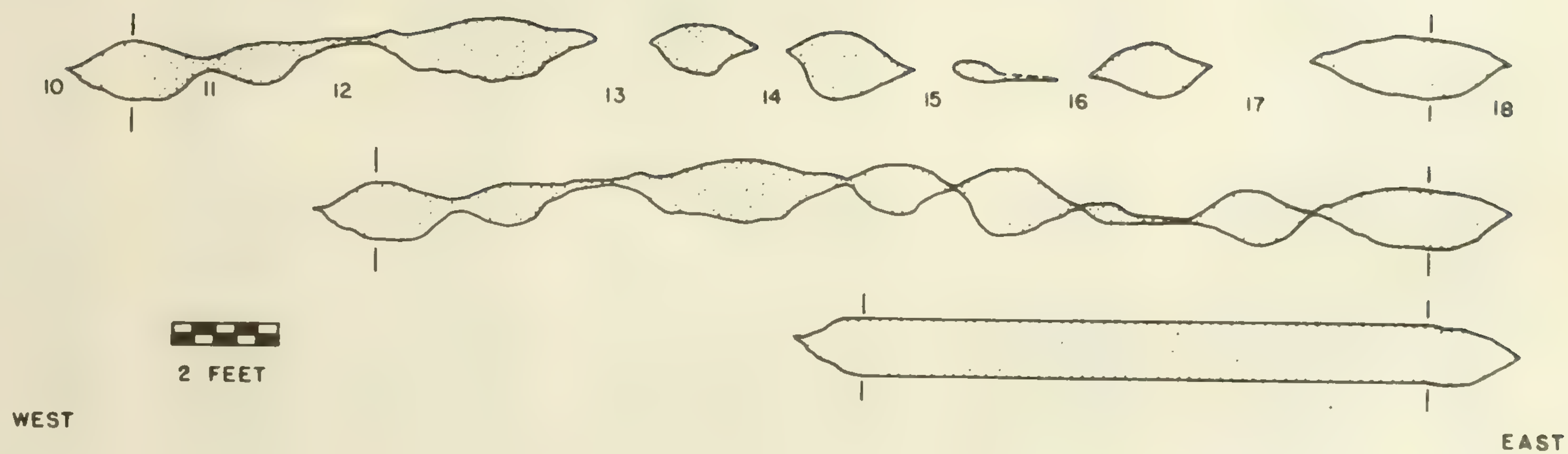


Figure 19. East-west transverse profile and restoration of boudin array 10-18, Locality 6.

locale characterized by repeated layers with similar physical properties and boundary conditions, spacing varies primarily as a function of boudin-layer and matrix-layer thickness variations.

A preliminary account of boudin mechanics has been given by Voight (1965); in that study an attempt was made (not without error) to theoretically account for such phenomena as spacing of boudins. Figure 20, taken from that study, includes data from this locality; the data have not been corrected for vein fillings. Portions of this work on boudin mechanics have since been revised but remain unpublished.

A distinctive east-to southeast-trending lineation is present within the plane of secondary foliation; these lineations are penetrative; they are not restricted to discrete surfaces and they do not simply represent ordinary slickensides produced by late-stage flexural fold deformation. Development of the lineation is syngenetic with secondary foliation, and hence with early folding and boudinage.

The mean orientations of neck lines (i.e., boudins) and penetrative lineations are approximately orthogonal at most localities (cf. Figures 21, 22), and this relationship is generally borne out by individual field observations where penetrative lineation and well-developed boudins occur in association. This relationship is interpreted to suggest that the penetrative lineations represent the direction of principal extension within the plane of flattening. Locally, distorted fossils can be found which seem to reinforce this view. Planispiral gastropods found at this locality were elongated parallel to penetrative lineations and perpendicular to boudin axes.

Return to automobiles and drive eastward to the Webster School. Continue across Route 30, on Route 73 in the direction of Brandon. Continue past Mrs. Selleck's General Store on Pleasant Brook, where a limited but fine selection of Vermont cheese was always available for a hungry geologist, to School No. 1. Distance from Webster School about 1.2 miles.

Locality 7. Parautochthon and Autochthon: Bald Hill, Stony Hill, Miller Hill, and Vicinity - Beginning at the western edge of Figures 23 and 24, reference is made to the nearly continuous bluff of white marble, which can be traced from Route 30 (south of the Webster School) to a ridge of glacial debris 400 yards west of School No. 1*(cf. Figures 8, 23). The marble over-

* The following may be of assistance to the reader in understanding the descriptions of the Bald Hill area, Figures 23 and 24. Reference: 7 1/2' U.S.G.S. Sudbury, Vermont, quadrangle. The Bald Hill area is traversed by Route 73, passing from the Seth Warner Memorial Highway at the Webster School, to Brandon. South of Route 73 is Stony Hill (800 ft. contour); north of Route 73 is Hill 641, and Bald Hill (713 ft. elev.); east of Stony Hill and Route 73 is Miller Hill.

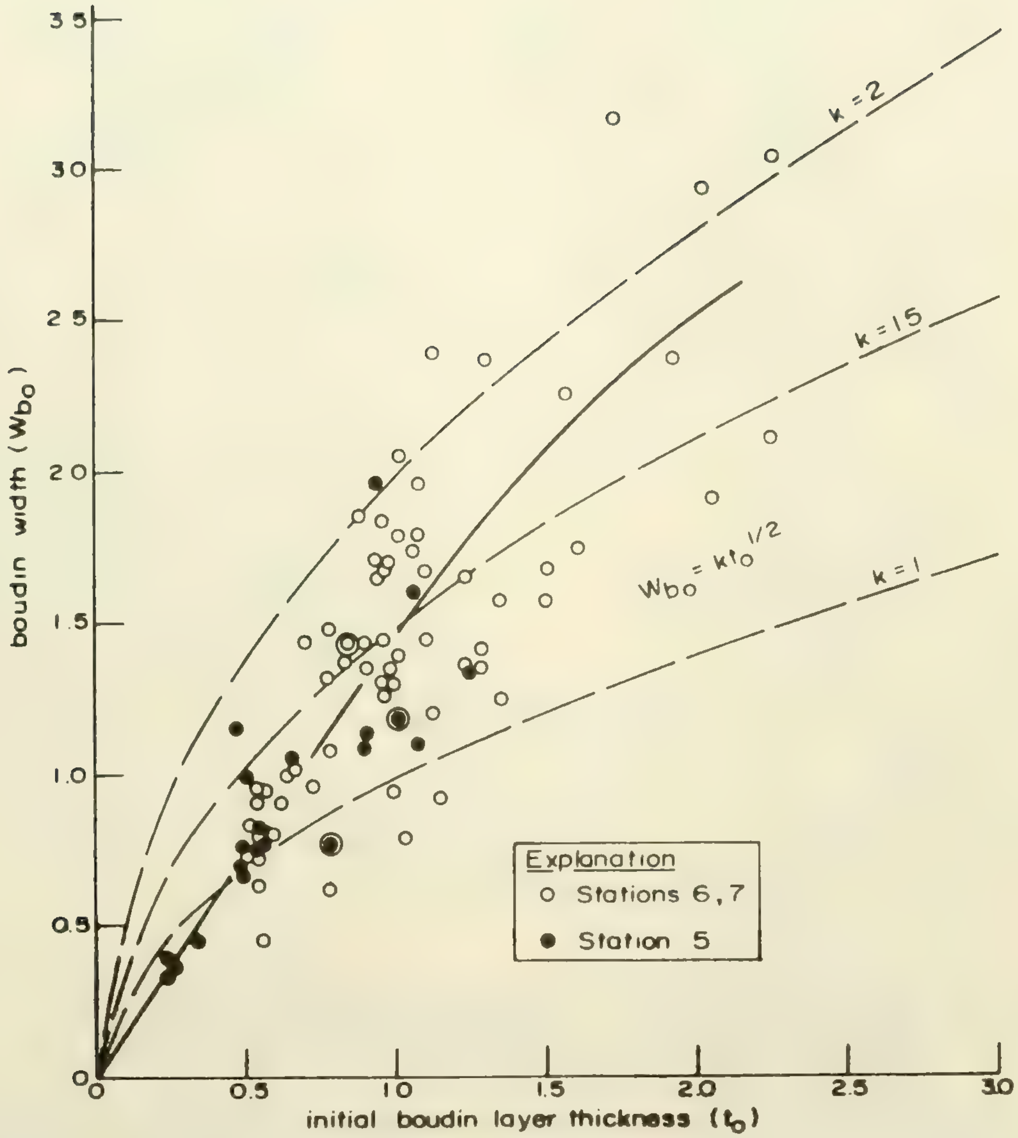


Figure 20. Initial (restored) boudin width versus initial boudin layer thickness. Station 7 of this figure corresponds to Excursion Locality 6.

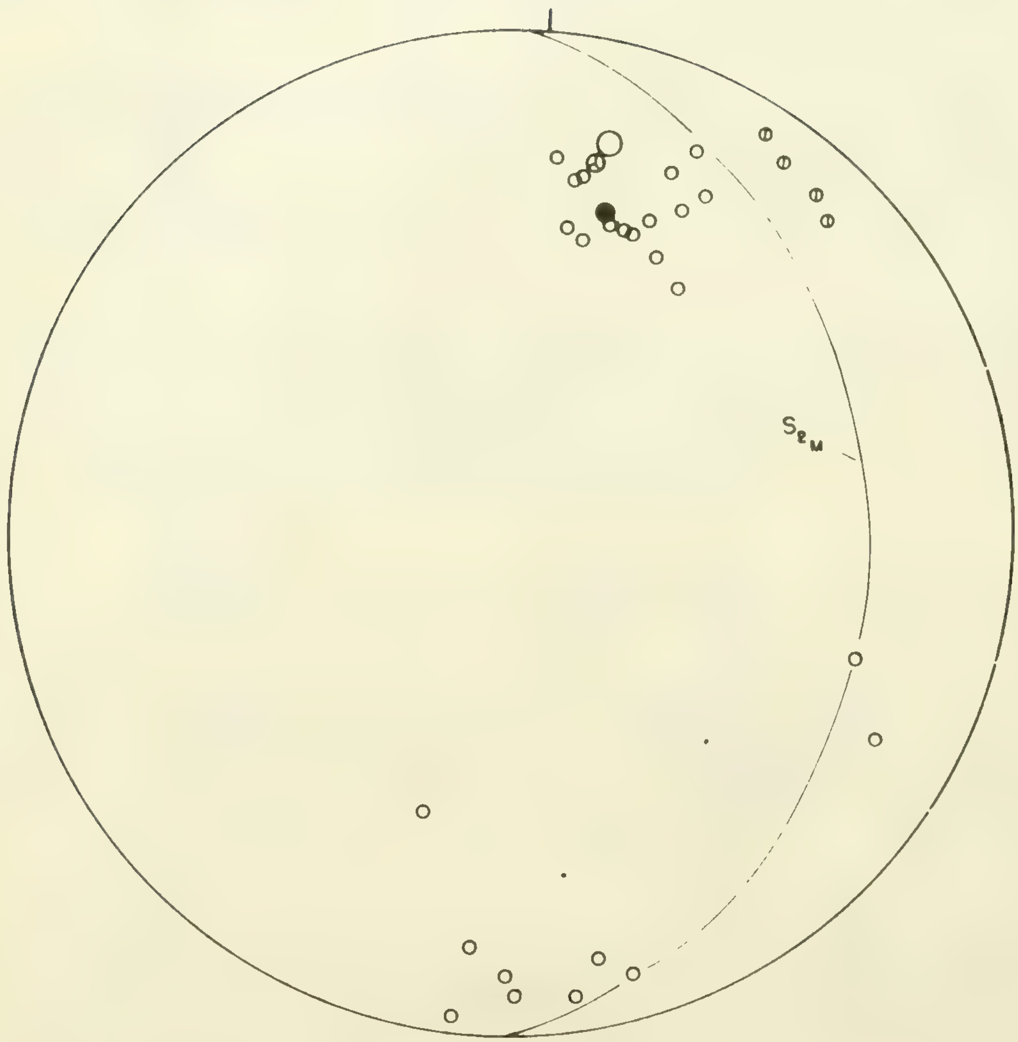


Figure 21. Synoptic diagram of boudin maxims, Sudbury nappe; Station 8 data (Miller Hill klippe) given for original (split circle) and rotated (circled dots) positions.

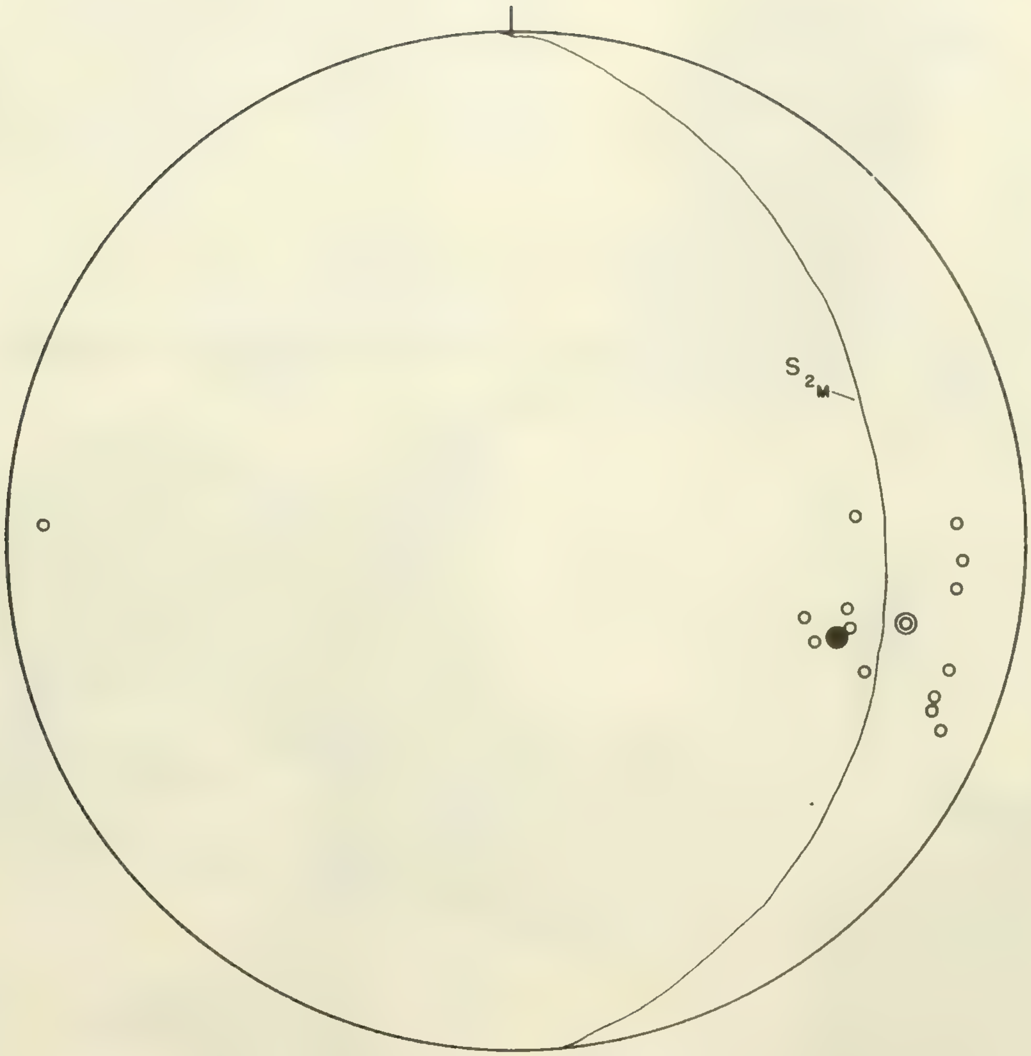
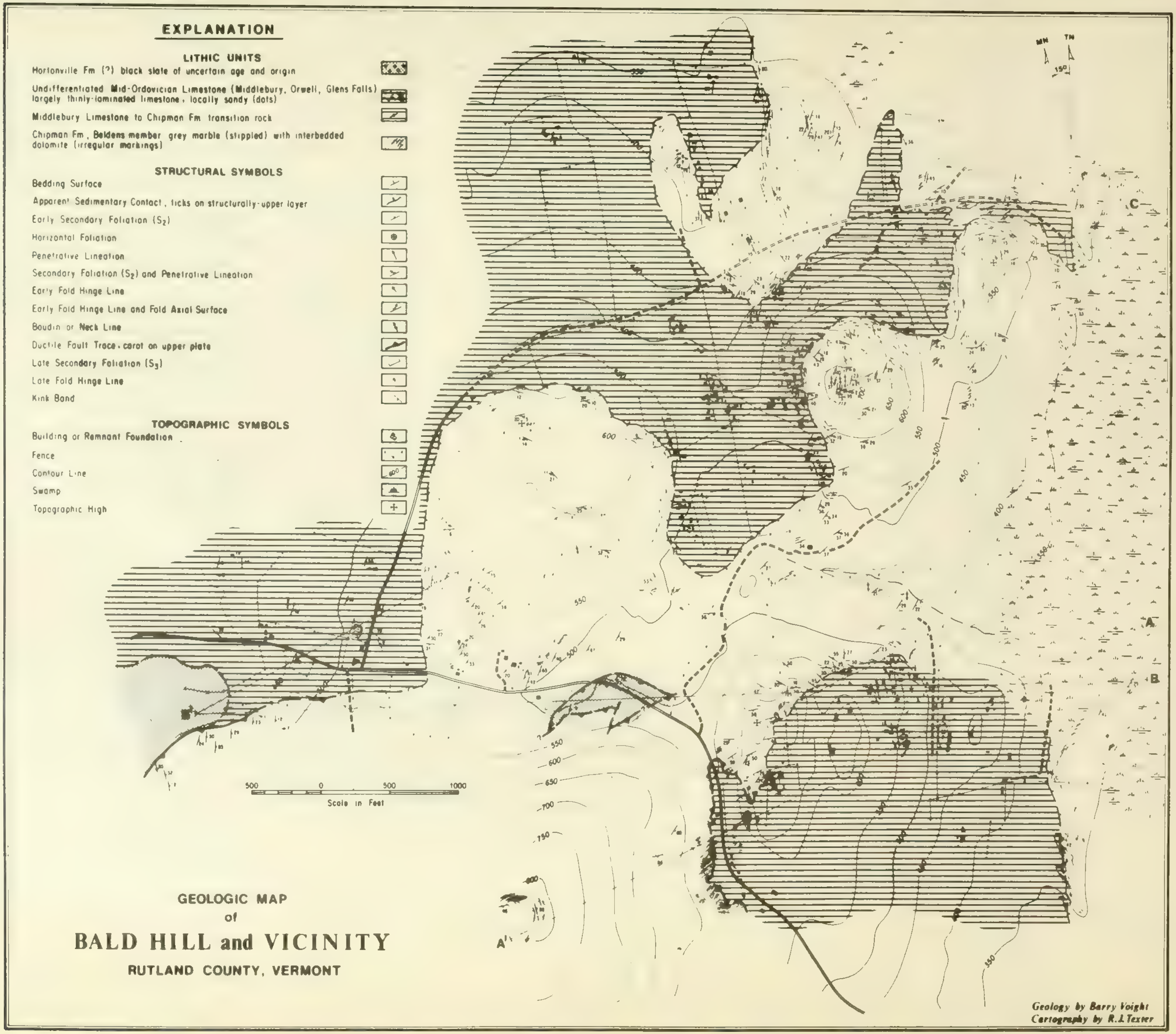


Figure 22. Synoptic diagram of penetrative lineation maxima, Sudbury nappe.



Geologic map of Bald Hill and vicinity. Geologic map patterns emphasized.

Figure 24. Geologic map of Bald Hill and vicinity; Structural data emphasized.

lies and truncates the contact of slate and mid-Ordovician limestones in the synclinorium core. This is the Sudbury thrust of Cady (1945, p. 570). The thrust contact is exposed locally; a 20° dip discordance exists between the thrust and foliation.

To the east, structural relationships seem more complex. In western ledges of Hill 641, Beldens marble appears to grade stratigraphically-upward into Middlebury limestone across a contact sub-parallel to dolomite bedding and secondary foliation. A typical sequence is as follows: beginning with the oldest, (1) white marble with massive dolostone beds seems to grade into (2) blue-grey calcite marble with massive dolostone, to (3) blue-grey marble, to (4) sugary-textured slightly argillaceous calcisiltite, often dolomitic, to (5) argillicalcilutite, typical of Middlebury limestone. The section is inverted at Hill 641; the contact can be traced to Brandon Swamp, and appears to be essentially stratigraphic although subjected to large strains; there seems to be no evidence for an unconformity as suggested by Zen (1961, p. 321).

The summits of Bald Hill and Hill 641 are interpreted as containing the overturned limb of a recumbent anticline with a core of Beldens marble. Amplitude exceeds one mile, and is of course much greater if this structure is continuous with the main body of the Sudbury nappe. Several smaller structures exposed on the north flank of Bald Hill, possibly recumbent anticlines, or nappes (with basal shear zones), have similar characteristics of style (Figures 23, 25). The Sudbury nappe may be in fact completely detached from subjacent structure, for field evidence is inconclusive on this point; nonetheless a genetic relationship amongst these structures seems most likely, and they are regarded as parasitic "digitations" on the overturned limb of the Middlebury synclinorium, here considered as a complex early fold.

On the north slope of Stony Hill, a sliver of slate (of unknown age and origin) occurs athwart Route 73. The contact is well exposed about 200 feet south of the road, with foliation within both slate and marble approximately parallel to the southeast-dipping contact. Zen (1961, plate 1) shows this slate as a continuous unit connected to the synclinorium core. Because of a small gap in outcrops, evidence to the contrary is not conclusive, but it favors no connection. The sliver is interpreted as either a thrust sliver, bounded top and bottom by "ductile" faults masked in secondary foliation.

Further to the east, detailed relationships between marble and limestone have been literally uncovered by hand-dug excavations along the contact zone. Structural discordance has been observed at a small klippe on Miller Hill, in ledges 200 ft. north of the klippe, at the eastern foot of Miller Hill near Brandon swamp, and at the foot of Stony Hill west of Route 73. Some of the contact relationships are schematically shown in Figure 26, in which figure the effects of subsequent folding have been ignored. Near the

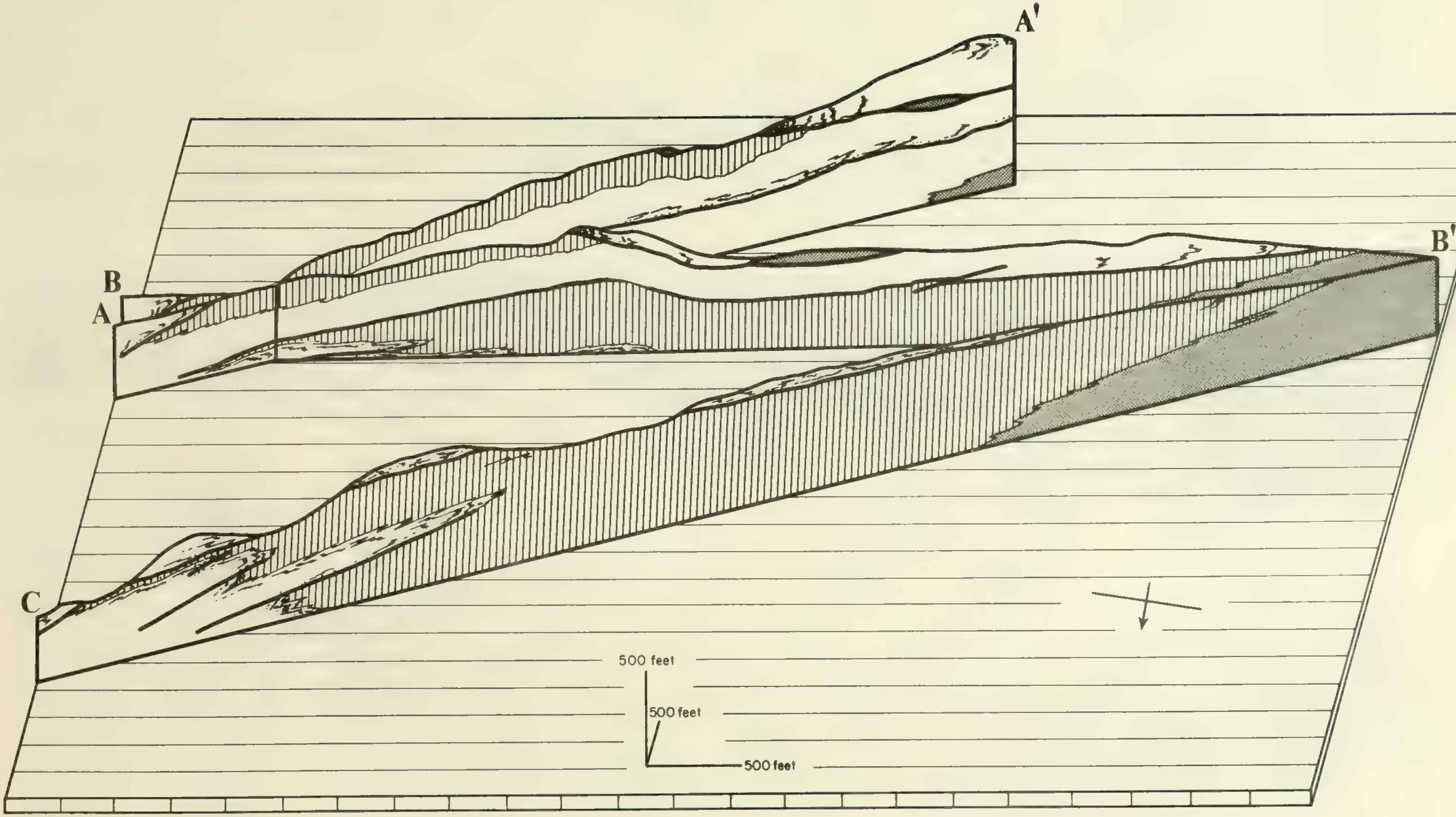


Figure 25. Cross sections of Bald Hill and vicinity (cf. Fig. 23).

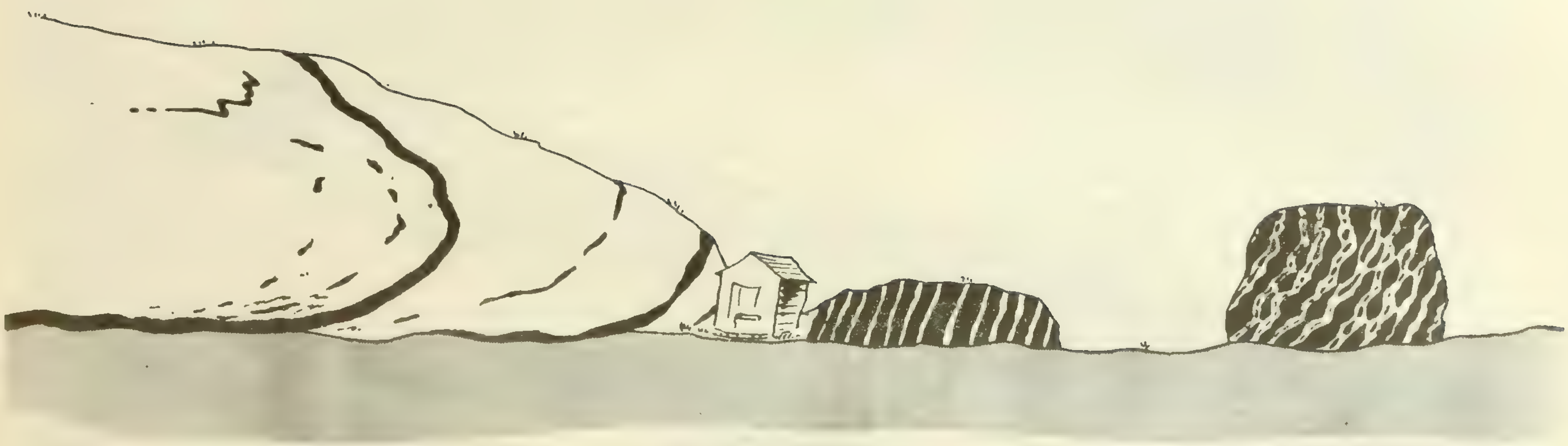


Figure 26. Schematic structural relationships in the vicinity of Miller Hill; black is dolostone, white is calcite marble (Beldens member, Chipman formation); grey is Middlebury limestone.

"Spring House"(shown in the figure) west of Route 73, a subhorizontal fault contact sharply truncates steeply-dipping bedding (dolostone and marble) within the Beldens marble. Similarly, across Route 73, at the klippe, the fault truncates steeply-dipping, northeast-striking dolostone and marble layers. The rocks have been severely strained, as evidenced by inter-meshed boudins in the dolostone, superbly shown on the west face of the klippe*; much marble has been squeezed out in conjunction with boudinage evolution, which is presumed to be approximately synchronous with the large-scale low-angle thrust faulting. These outcrops seem to represent a dissected thrust plate -- the Miller Hill thrust (Voight, 1965).

The fault contact relationships vary as a function of rock type. South of the "spring house", in a bluff where more than fifty feet of continuous exposure occurs along the contact, a massive dolostone layer about one foot thick directly overlies the contact (Figure 26). Secondary foliation within marble and subjacent limestone essentially is parallel to the contact, which contact locally varies in attitude from vertical to sub-horizontal, presumably on account of subsequent folding. Despite this "apparent conformity" at the contact, rock flowage has been intense, as evidenced by broken fragments of dolomite within the marble; the dolomite beds have changed systematically in attitude, from an east-northeast strike, steeply-dipping attitude to virtual parallelism with the contact, a consequence of a "drag" effect associated with the Miller Hill thrust. This thrust appears to be a discontinuity within a mass undergoing large strains, i.e., presumably a discontinuity in terms of displacement, strain, and velocity. Locally concordance or discordance may be apparent, depending upon the rheology (e.g., ductility) of the bordering rock. Within the environment of deformation associated with the Miller Hill and Sudbury thrusts, the deformational mode for dolomite was by fracture; under identical conditions, calcite readily flowed. Hence where dolomite formed the predominant rock type, brittle fracture and discordant relations are observed; where marble predominates, the contact relationships can seem to be concordant. Incidentally, it should be mentioned that there is some danger of misidentifying secondary foliation as bedding, particularly as thin-bedded limestone. Decisions concerning concordance or discordance, which concern data on foliation rather than bedding, are to be interpreted cautiously.

On the northern slope of Miller Hill, near the 550-foot contour, the marble unit appears to underlie the limestone. Thus a structural inversion of the marble-limestone contact apparently takes place at Miller Hill. The contact is definitely a fault contact in the vicinity of Route 73, but no definitive evidence

* Boudin orientation is shown in Figure 24; by unrolling axes in small-circle paths about the "late" fold axis to account for rotation of the fault surface, the pre-folding orientations were reconstructed. An initial horizontal attitude of the folded surface was presumed.

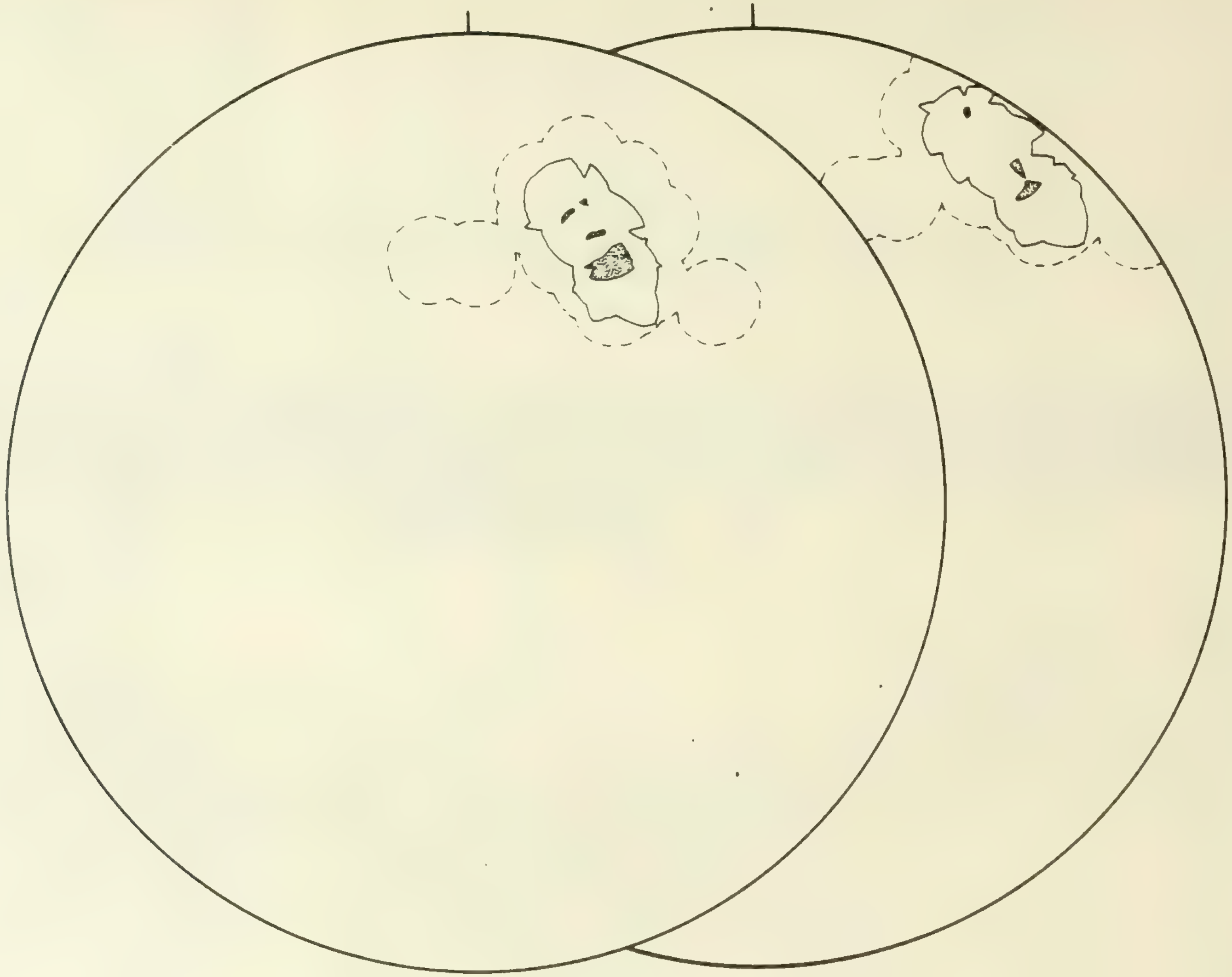


Figure 27. Distribution of neck lines in the klippe at Miller Hill; rotated (left) and original (right) data (cf. Fig. 21).

favorable to either a concordant or discordant hypothesis was observed where the marble underlies limestone.

Amongst other possibilities, the "inversion" could represent a folded thrust, or a thrust plate of marble overlying a sequence of marble and (younger) limestone in "normal" order.

Part of the interpretation problem for the entire Bald Hill area arises from the fact that thrust relationships are recognizable only where dissimilar rocks are in contact -- and sometimes not even under those conditions. No structural breaks are recognizable where marble rests upon marble, despite the fact that exposures are unusually good for this region, a locale of Alpine tectonics typically lacking alpine exposures. This is not to deny the existence of faults within the marble, only our ability to recognize them; these faults are principally ductile phenomena, masked by and associated with the development of secondary foliation.

A unique explanation satisfying all field relationships is not known to me. Two hypotheses are given here for purposes of discussion by excursion members (cf. Figure 28):

(A) double-nappe hypothesis; Miller Hill nappe overlies the Sudbury nappe;

(B) single-nappe hypothesis; Miller Hill and Sudbury thrusts are part of a single structural element.

Additional hypotheses or variants of the above may be suggested, none of which in detail have the attribute of simplicity. Ultimately, however, it should be noted that the "key" to understanding the structure at this locality hinges on two units, the slate north of Stony Hill, and the limestone at Miller Hill. A blank cross-section, approximately an extension of line A'-A on Figure 23, is provided for the excursion participants (Figure 29) in order to provide an opportunity for individual interpretation based on the evidence presented at this stop.

An overall view of the structure associated with "Sudbury nappe" is shown in Figure 30; the right side of this figure depicts the geology at Locality 7. and shows its assumed relationship to the structure of the Middlebury synclinorium and Taconic allochthon.

Return west on Route 73 to Route 30; drive north 14 miles to Middlebury, Vermont, and junction Route 7. Park on north side, Otter Creek; walk to below falls.

Interlude: This region marked a main route for the Algonquin and Iroquois; beginning on Lake Champlain (the Iroquois "Gate of the Country"), the route followed Otter Creek from Basin Harbor to its

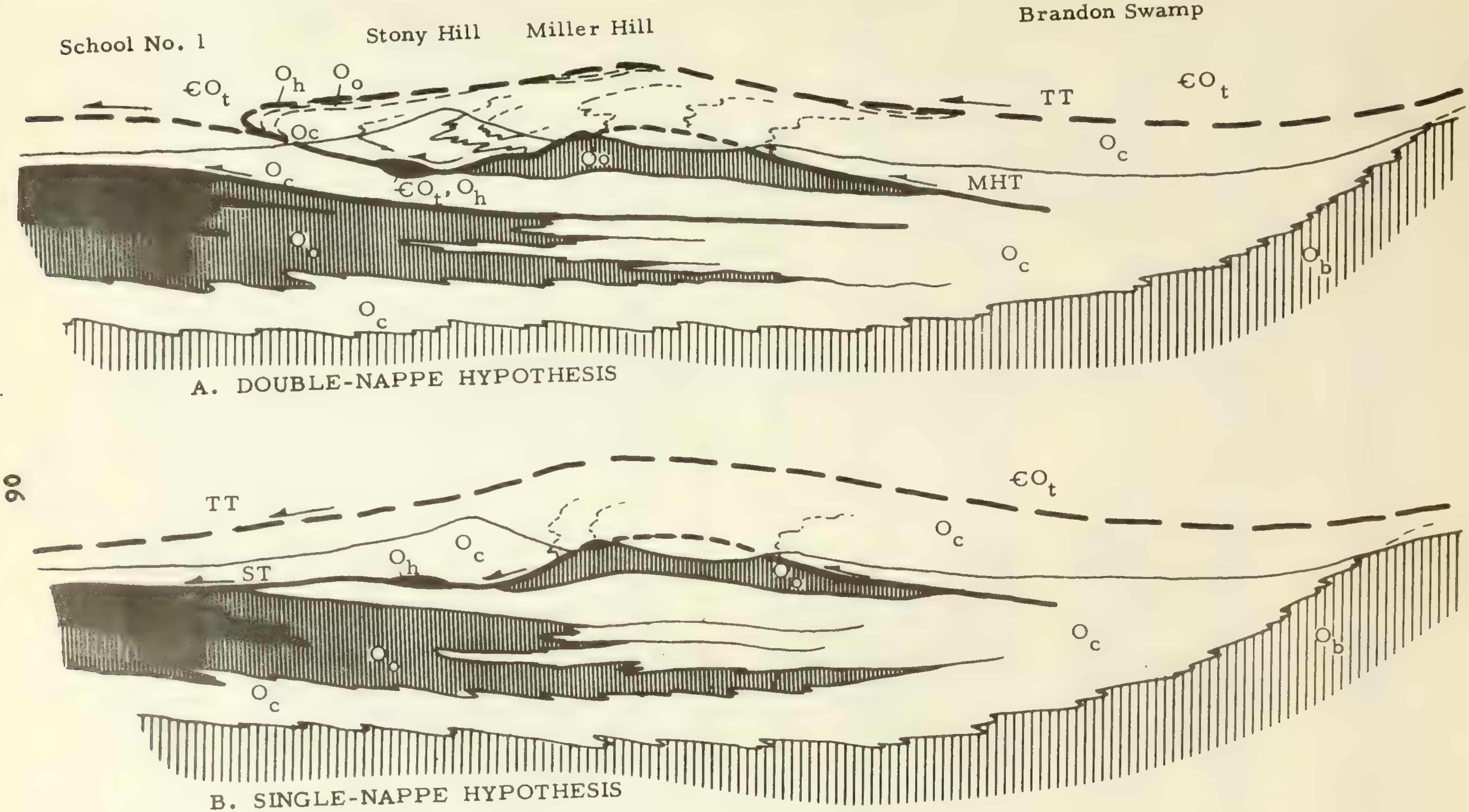


Figure 28. Schematic diagram of two structural interpretations for the Stony Hill-Miller Hill area.

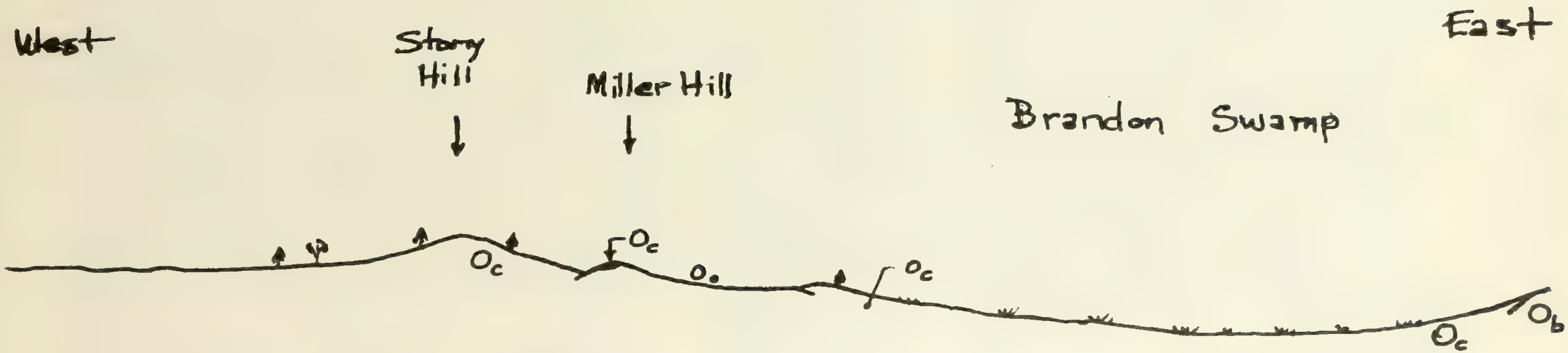


Figure 29. Topographic profile through Stony Hill-Miller Hill area. Geology to be filled in by excursion participants.

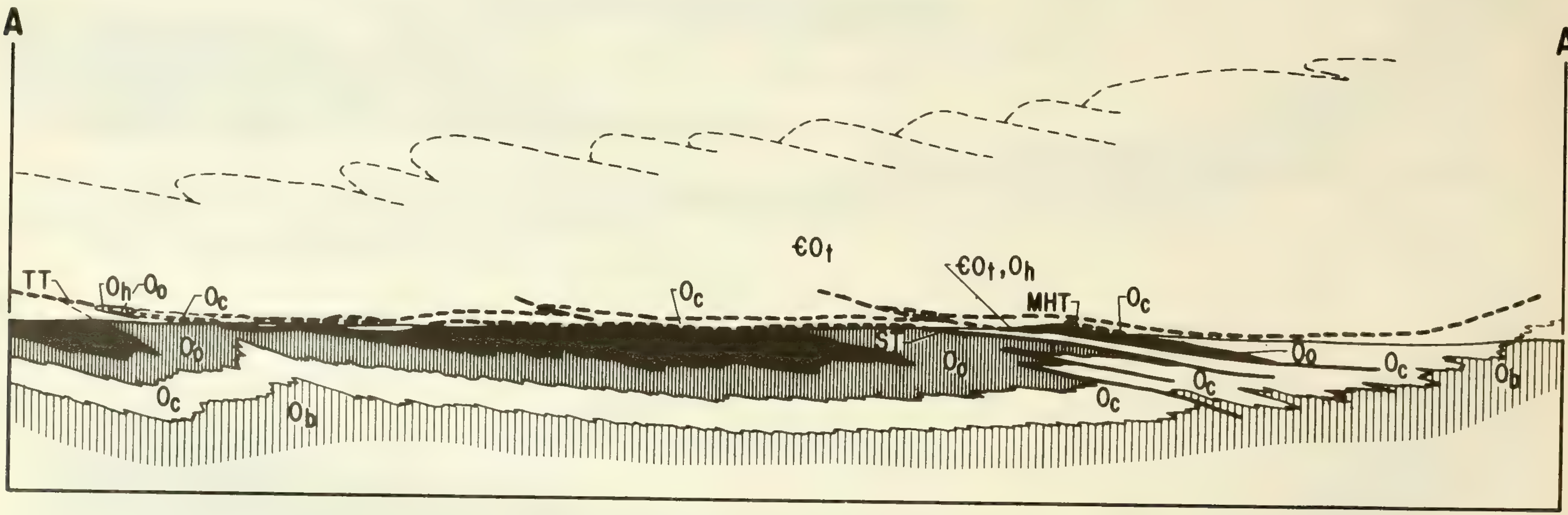


Figure 30. Geologic cross section of the Sudbury nappe area, showing its relationship to the superjacent Taconic allochthon and subjacent Middlebury synclinorium. Stratigraphic symbols: Ob, Bascom formation; Oc, Chipman formation; Oo, undifferentiated mid-Ordovician limestones; Oh, Hortonville formations; €Ot, undifferentiated "Taconic sequence" rocks. Structural symbols: TT, "Taconic thrust"; ST, Sudbury thrust; MHT, Miller Hill thrust.

headwaters. Here a portage followed, from the Vermont Valley across the Green Mountains to Weston, thence along the West River to the Connecticut River at Brattleboro, southbound to Connecticut. Another main route followed the Lemon Fair River (Locality 6.) from its headwaters to Lake Champlain; this route was used by Mohawks from the Hudson Valley. The name Bomoseen means "big pond with grassy banks" in Abnaki language; the lake of that name (Locality 3.) was used by Algonquin and Iroquois to catch fish, which were smoked and carried to their permanent winter lodges. Certain parts of the Otter Creek and Champlain valleys were previously occupied by peoples associated with the "Laurentian culture"; Laurentian sites in New York have been dated at 2500-3000 B.C. Little appears to be known of the ancient peoples, sometimes referred to as the "red paint" or "slate" culture; excavations of living sites had been carried out in Orwell by the Heye Foundation of the Museum of the American Indian, New York City, in 1933-35. A unique custom of the mourners was to place red hematite, symbolic of life, with their regarded dead.

Locality 8. East limb of the Middlebury Synclinorium; Middlebury Village - The critical Beldens-Middlebury contact has been traced northward from the Bald Hill vicinity in order to see if the style of deformation evident at Bald Hill was present along the entire eastern edge of the Middlebury synclinorium. Much of this contact is covered, but where exposures permit, the contact seems to dip moderately eastward, generally approximately parallel to secondary foliation.

As an example, a locality several hundred feet west of Middlebury Village will be examined. Beldens marble, containing superb exposures of north-trending boudins in thick massive dolostone beds, lies in inverted contact with Middlebury limestone on the north bank of Otter Creek below the falls (cf. Cady, 1945, Plate 4, Figure 3; Seely, 1910, p. 30, plate 39). Foliation parallels the contact, dipping eastward at 40°. Absence of a transitional zone and profound attenuation of dolostone above the contact (as evidenced by boudinage) suggest a thrust contact involving differential flow between the two rock units. The limestone below the thrust has very likely undergone strains on the same order as marble overlying the thrust, as suggested by the presence of foliation; however its magnitude cannot be estimated in the absence of structural elements such as boudinage.

Field observations show that the Beldens marble - Middlebury limestone contact is inverted throughout the east synclinorium limb (Voight, 1964b), and suggest that phenomena described at Bald Hill are not restricted to that locality. The observed relationships are compatible with the hypothesis that the sequence Bascom formation - Middlebury limestone is involved in (or comprises) a "root" zone characterized by differential, and locally discontinuous, flow (Figure 30). The "root" is taken to be that

zone from which a nappe arises, and as such, in the present instance, represents a zone of detachment between the nappes and less extremely deformed foreland rocks of the synclinorium east limb.

Take Route 7 toward Burlington, 33 miles to the north.

Interlude: It's fun to watch the way they're made,
They wasn't built by grafters;
The cords, the uprights, oak-hewed pins,
The ridge pole and the rafters;
An iron bridge turns rusty red
A concrete bridge gets sooty;
Give me a good old covered bridge
For business, love, or beauty.

Locality 9. The Synclinorium core: Ledge Creek section. (To be visited if time permits.) - Inversion of contacts extends to the synclinorium core, and can be demonstrated, e.g., at Ledge Creek, 3 miles northwest of Middlebury Village. Here, with about a quarter-mile of continuous exposure, the Glens Falls limestone-Hortonville slate section can be studied from the west to the east limb of the Middlebury synclinorium. East-dipping early foliation is consistent with an axial-plane relationship to the overturned section, which seems precisely as shown by Augustus Wing (see Cady, 1945, Figure 2) in a section drawn approximately through this same locality. The synclinorium is not merely a late, relatively open fold structure, the concept apparently accepted by many workers of recent vintage, but has an origin dating back to the development of isoclinal fold structures, associated foliations, and ductile faults such as the Miller Hill thrust.

Interlude: Reader, what I have here written, is not a Fiction, Flam, Whim, or any sinister Design, either to impose upon the Ignorant, or Credulous, or to curry Favor with the Rich and Mighty, but in meer Pity and pure Compassion to the Numbers of Poor Labouring Men, Women, and Children in (New) England, half sterv'd, visible in their meagre looks, that are continually wandering up and down looking for Employment without finding any, who here need not lie idle a moment, nor want Encouragement or Reward for their Work, much less Vagabond or Drone about it. Here are no Beggars to be seen (it is a Shame and Disgrace to the State that there are so many in England) nor indeed have any here the least Occasion or Temptation to take up that Scandalous Lazy Life.

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Plate 1. Polished cross-section across the "neck" of a dolostone boudin from Locality 6. Dolostone is dark grey, calcite and quartz veins are white, marble (matrix) is light grey. The fragmentation and brittle fracture of dolostone within the neck region is well shown; under the "same" conditions, the surrounding calcite marble layers deformed by flowage.

Trip B-4

THE CHAMPLAIN THRUST AND RELATED FEATURES NEAR MIDDLEBURY, VERMONT

by

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This field trip will review preliminary results from investigations of the Champlain thrust, Middlebury synclinorium, and Green Mountain anticlinorium near the latitude of Middlebury by senior geology majors at Middlebury College. During the past seven years 20 senior theses have been completed; seven of these provide a nearly continuous geologic map of the Champlain thrust belt between Vergennes and Route 125 west of Middlebury at a scale of 500 feet to the inch. Other theses have included regional and local gravity studies, geologic mapping of critical localities in the Middlebury synclinorium and Green Mountain anticlinorium, petrologic studies of greenstone and ultramafic bodies east of the anticlinorium, sedimentological studies of lower Paleozoic rocks, and mapping and petrologic studies of Mesozoic igneous rocks. These efforts have built upon earlier studies in west-central Vermont by Cady (1945), Welby (1961), Osberg (1952), and the unfortunately unpublished work of Crosby (1963). Although the field trip will concentrate on the Champlain thrust west of Middlebury (Figure 1), the regional tectonic setting of the thrust is briefly discussed here as background for participants.

TECTONIC SETTING

At the latitude of Middlebury four distinct tectonic provinces are from west to east: the Adirondack massif, the Lake Champlain lowland, the Middlebury synclinorium bounded on its west side by the Champlain thrust belt, and the Green Mountain anticlinorium. The tectonic significance of the Champlain thrust must be sought in the nature of these provinces and their boundaries.

In the Lake Champlain lowland (Welby, 1961) a relatively undeformed 5,000 foot sequence of Upper Cambrian through Middle Ordovician clastic and mainly carbonate shelf assemblage rocks rest with profound unconformity on a crystalline Precambrian basement (Figure 2). The Precambrian is extensively exposed in the Adirondack massif. The boundary between the Adirondacks and the lowland is a complex of fault blocks, down-faulted to the east, and structural relief on the Precambrian basement is at least 5,000 feet.

To the east of the Champlain lowland an Eocambrian to Middle Ordovician clastic and mainly carbonate shelf assemblage nearly 10,000 feet thick presumably rests on Precambrian basement (Figure

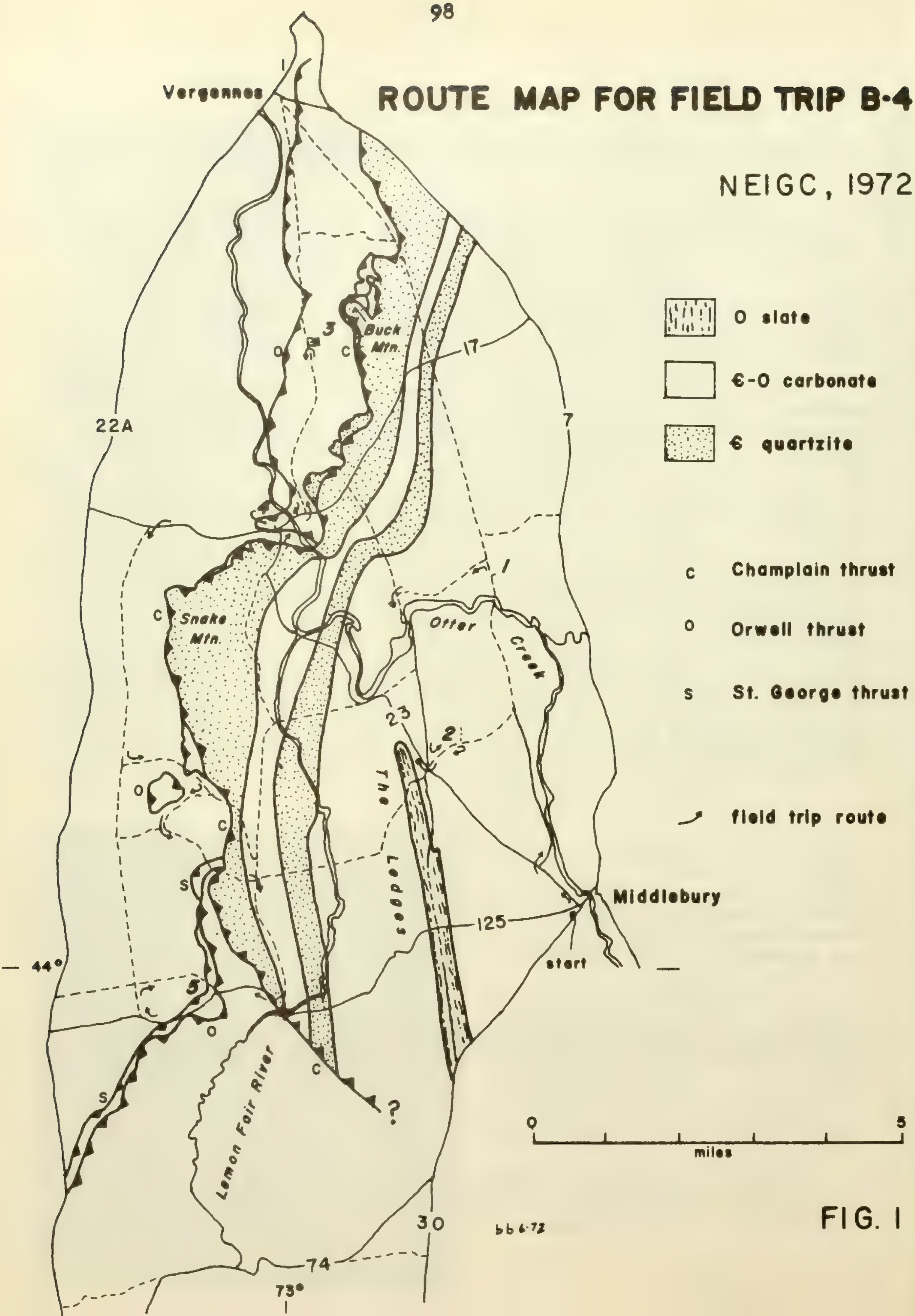
ROUTE MAP FOR FIELD TRIP B-4
NEIGC, 1972
O slate
Є-O carbonate
Є quartzite
c Champlain thrust
o Orwell thrust
s St. George thrust
field trip route
Vergennes
Buck Mtn.
Snake Mtn.
Otter Creek
The Ledges
Lemon Fair River
Middlebury
start
22A
17
7
23
125
30
74
44°
73°
0
5
miles
bb 6-72

FIG. 1

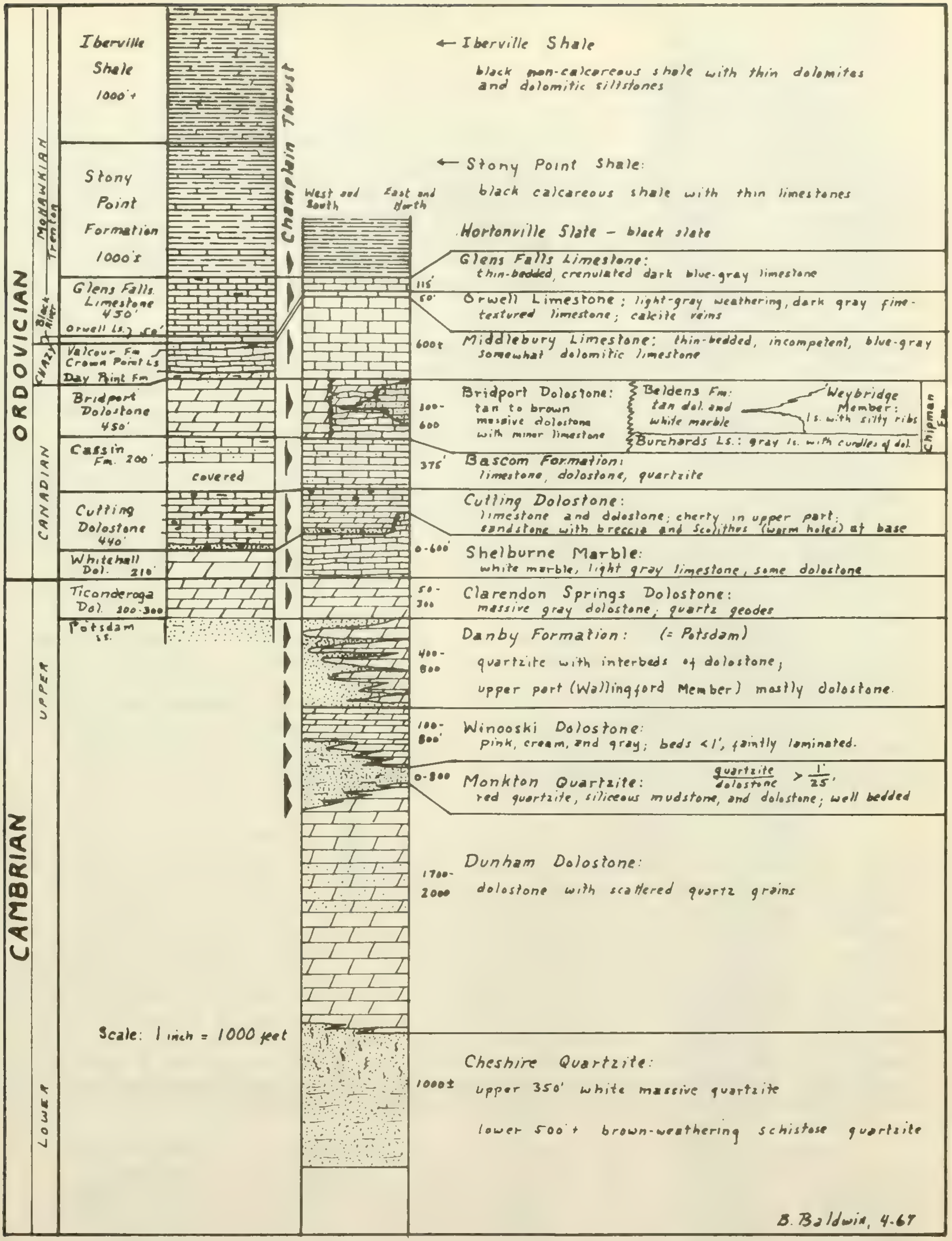

FIG. 2. Sequence of strata near Middlebury, Vt.

Compiled from Cady, 1945; Kay and Cady, 1947; Cady and Zen, 1960; Welby, 1961.

2). This sequence is rather intensely deformed into the south-plunging, westward inclined, Middlebury synclinorium (Cady, 1945). The two disparate early Paleozoic shelf assemblages of the Lake Champlain lowland and the Middlebury synclinorium (Figure 2) are separated by the Champlain thrust belt. The thrust belt is a series of east-dipping, low-angle faults which can be traced from at least southwestern Vermont and the adjacent New York northward into Canada (Cady, 1969).

The Green Mountain anticlinorium (Cady, 1945) rises sharply just east of the Middlebury synclinorium and is a west-vergent, doubly plunging, complex anticlinorium with an exposed core of Precambrian Mount Holly basement (Figure 3). Structural relief on the Precambrian unconformity between the floor of the Middlebury synclinorium and the crest of the anticlinorium is at least 3 miles, and probably as much as 6 miles, in an east-west distance of about 8 miles (Powell, 1969; Tennyson, 1970). The boundary between the synclinorium and the anticlinorium at the surface is locally marked by east-dipping thrusts, but is mainly a descending cascade of west-vergent folds (Osberg, 1952; Tennyson, 1970) termed the Green Mountain front.

The Green Mountain front is a major stratigraphic as well as tectonic boundary marking an abrupt facies change in Eocambrian to Ordovician rocks from the mainly carbonate miogeoclinal shelf assemblage on the west to "eugeoclinal" graywacke assemblage on the east. The Taconic "klippe", also of the eastern eugeoclinal facies, now lies athwart the Middlebury synclinorium on top of shelf rocks of the same age to the south of Middlebury, but has been removed by erosion northward.

With the exception, presumably, of the Chester dome, no bona fide "Yankee" Precambrian cratonic basement is known east of the Green Mountain anticlinorium. Significantly a belt of serpentinized dunites (Beyer, 1972) and meta-greenstones (Crocker, 1972; Doolan and others, in preparation) lie just east of the anticlinorium embedded in meta-graywackes of the early Paleozoic "eugeoclinal" assemblage. These rocks certainly mark a most significant tectonic boundary. It would appear, thus, the entire lower Paleozoic North American continental margin assemblage is exposed in a belt now less than 50 miles wide. Putting it another way, in the context of current plate tectonic theory, once one takes a single step east of the Green Mountain anticlinorium everything to the Bay of Maine is of suspect geo-political allegiance.

An unconformity of late Middle Ordovician age seen at one place or another in most of the region, including the Taconic "klippe", separates the apparently west-derived shelf and "eugeoclinal" assemblage from an apparently east-derived detrital-shale assemblage. The critical overlying rocks are the Hortonville Formation in the Middlebury synclinorium, the Pawlet Formation in the Taconic "klippe" (Zen, 1962), and the Moretown Formation east

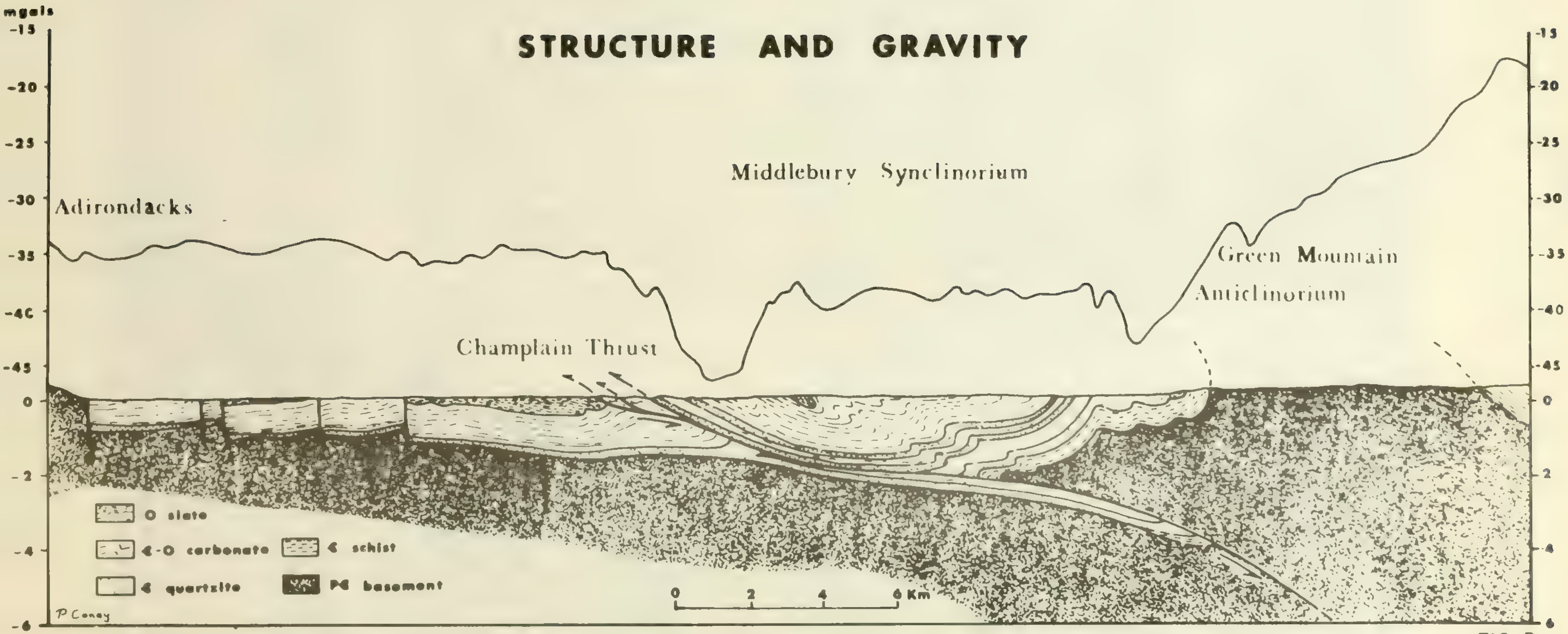
STRUCTURE AND GRAVITY
mgals
Adirondacks
Middlebury Synclinorium
Green Mountain
Anticlinorium
Champlain Thrust
O slate
Є-O carbonate
Є schist
Є quartzite
PЄ basement
0 2 4 6 Km
Km
P Coney

FIG. 3

of the Green Mountain anticlinorium. Silurian-Devonian detrital rocks mask all prior tectonic relationships east of the Moretown Formation to the Ordovician Oliverian volcanic arc along the Vermont-New Hampshire border (Rodgers, 1970), the presumed source of the detrital-shale flood. Very preliminary studies by Sedgwick (1972) suggest the Moretown Formation has a distinctly different heavy mineral assemblage compared to all older west-derived shelf and "eugeoclinal" rocks sampled to date.

The Champlain and related thrusts and the folds in the Middlebury synclinorium involve rocks as young as late Middle Ordovician. The east side of the Green Mountain anticlinorium is often argued to involve Silurian-Devonian rocks as well (Cady, 1968). If all these major structures have any genetic relationship to one another, as is generally assumed, much deformation is as young as Acadian (Middle Devonian) at least. The emplacement of the Taconic "klippe" is well documented as Taconic (Middle Ordovician) and numerous ductile folds and minor structures beneath and adjacent to the "klippe" in synclinorium rocks are correlated with this emplacement (Crosby, 1963). All these "Taconic" structures, including the "klippe" (Crosby, 1963; Johnson, 1970) are redeformed by younger more "brittle" deformation thought to be Acadian (Crosby, 1963). The Middlebury region, then, has suffered at least two deformations. To what extent these phases were discrete events or a single continuum is yet to be resolved.

THE CHAMPLAIN AND RELATED THRUSTS

General Statement. The Champlain and related faults form a belt of east-dipping, low-angle thrusts which separate the Middlebury synclinorium on the east from the Lake Champlain lowland to the west. The belt of thrusts brings up resistant rocks, such as Cambrian Monkton Formation quartzite, which have produced a line of hills and ridges. Snake and Buck Mountains are the prominent ridges at the latitude of Middlebury. The system of faults forms a tightly packed series of slices exposed in a belt seldom more than 2 miles wide. The Champlain thrust is the most easterly of the faults while the other faults lie just west of, and structurally below, the Champlain thrust. For regional stratigraphic detail the reader is referred to Cady (1945) and Welby (1961), and to Figure 2.

Champlain Thrust. The Champlain thrust (Stop 3) enters the area at the north end of Buck Mountain and extends southward for at least 15 miles to Route 125 west of Middlebury. The fault can be easily traced northward into Canada, but its fate south of Route 125 is still in question (Stop 5). Cady (1945) and the Centennial Geologic Map of Vermont (Doll and others, 1961) terminate it in the very poorly exposed south-plunging anticline just west of Cornwall several miles south of Route 125. Over almost the entire trace west of Middlebury, Cambrian Monkton Formation is

thrust over highly deformed Ordovician carbonate and shale.

The thrust plane lies within several hundred feet of the base of the Monkton Formation and only at the south end of Buck Mountain (Cady, 1945; Welby, 1961; Egan, 1968) does it bite lower into several tens of feet of Dunham Dolostone. The Monkton Formation on the upper plate generally dips gently eastward into the Middlebury synclinorium forming a prominent dip slope on the east sides of Buck and Snake Mountains. Near the fault trace, however, local imbrications and folds are evident in Monkton Formation layers. The fault plane is only rarely exposed, but at several points on Buck and Snake Mountains dips between 7 and 25 degrees eastward are seen.

The trace of the fault wanders in topography and has right en echelon offsets from the north end of Buck Mountain south to Snake Mountain. South of the summit of Snake Mountain it trends southward then angles southeasterly until lost beneath glacial cover south of Route 125. The en echelon offsets result in a series of salients and re-entrants. Structure contouring on the fault surface (Westervelt, 1967), gravity studies and geologic mapping (Davidson, 1970) suggest the salients are shallow downwarps and the re-entrants are mainly up-warps. Thus, the marked offsets are apparently primarily due to topographic expression of undulations in the fault plane rather than numerous cross-faults as shown on earlier maps (Welby, 1961; Doll and others, 1961). The southern "termination" near Route 125, where the trace angles southeastward, is marked by a southward structural plunge of the Monkton Formation before it and the fault trace are buried by glacial cover. If the fault continues southward it must climb up-section to place higher units on the sole. If the fault indeed terminates near Cornwall then the plunge of the Monkton is probably due to rapid decrease in dip separation into the anticlinal core (Smith, 1972).

Stratigraphic separation on the fault reaches nearly 5,000 feet. If the shallow dips of the fault surface are projected eastward beneath the Middlebury synclinorium, and reasonable reconstructions of geometry are made on the upper and lower plate, the dip separation is well over 10,000 feet. Depending on how far the Monkton Formation extended westward from its present exposures along the fault trace, the dip separation could be much more. If the fault terminates west of Cornwall this dip separation must decrease rapidly to zero, but where last seen Monkton Formation must lie over at least Bridport Formation giving a stratigraphic separation of about 3,000 feet.

Related Thrusts West of the Champlain Thrust. The lower plate of the Champlain thrust is made up of a series of thrust slices from north of Buck Mountain to its apparent termination near Route 125. South of Route 125 the lower thrusts angle to the southwest away from the last seen southeasterly trend of the Champlain thrust.

Rocks within the slices are generally intensely deformed into tight west-vergent folds.

At the south, near Route 125 (Stop 5), two thrusts are identified. The most westerly is the St. George thrust (Cashman, P., 1972; Cashman, S., 1972; Lyman, 1972) followed eastward by the Orwell thrust (Cady, 1945; Welby, 1961). The St. George thrust generally places Glens Falls Formation over Middle Ordovician shales whereas the Orwell thrust generally places Bridport Formation over Glens Falls Formation. Mapping at the north and south ends of Snake Mountain suggests the thrusts become younger from west to east since the St. George thrust is truncated by the Orwell thrust and the Champlain thrust truncates both the Orwell and St. George thrusts (Cashman, P., 1972; Cashman, P. and others, 1972; Cashman, S., 1972; Lyman, 1972). Dips on the fault planes of the two older thrusts vary. Of the two the St. George thrust often appears the steeper, and in several places it must exceed 30°E. Dips on the Orwell thrust are horizontal at the Crane School "klippe", or salient (Cashman, S., 1972), but elsewhere may be close to 10°E. At the south end of Snake Mountain both thrusts are truncated by the Champlain thrust placing Monkton Formation directly on Middle Ordovician shale north of the truncation. The Orwell thrust reappears from beneath the Champlain thrust on the north side of Snake Mountain and has been mapped northward to Buck Mountain. At Buck Mountain, particularly in the re-entrant north of the main summit, several small thrust slices lie between the Orwell and Champlain thrusts superimposing Bridport and Crown Point Formations in imbricate slices (Westervelt, 1967).

Although stratigraphic separations on the older thrusts are understandably less than those found on the Champlain thrust, dip separation on the Orwell thrust could be equally as large as that calculated for the Champlain thrust. In reality the soles of all these thrusts probably converge at depth beneath the Middlebury synclinorium.

Folds and Minor Structures. Most folds found directly below the Champlain thrust (Stop 3) are tight flexural-flow or passive folds usually inclined or overturned in westerly directions. A well developed axial surface "crenulation" or "fracture" cleavage is common, particularly in Crown Point limestones (Stop 3). Directly above the Orwell thrust, particularly in Bridport dolostones and limestones, fold patterns are more complex (Stop 4). Trends of folds are generally northward, but locally axial traces sweep into more latitudinal trends. In many places the folded bedding, particularly just above the Orwell thrust, is truncated by thrusting and the folds seem to have developed before as well as during faulting. At no place, however, has the axial surface crenulation cleavage of the folds below the Champlain thrust been found to be deformed by any penetrative later cleavage. On the other hand, numerous overturned to recumbent folds with variable trends are found in Ordovician shales below the thrust belt. Some of these

have a well developed axial surface cleavage, but one large fold on the west face of Snake Mountain where the Champlain thrust lies directly on the shale, has a north-trending cleavage unrelated to fold geometry (Lyman, 1972). Crosby (1963) reports evidence of two deformations south of Route 125 in Bridport Formation on the upper plate of the Orwell thrust. In general, however, evidence for two "distinct" deformations so characteristic of the Middlebury synclinorium to the east of the thrust belt is not so obvious in and west of the thrust belt.

GRAVITY DATA

About 500 gravity stations have been made in the vicinity of Middlebury. The results thus far are very preliminary. Two east-west profiles by Powell (1969; Powell and Coney, 1969) between the Adirondack Mountains and the east flank of the Green Mountain anticlinorium show a sharp, asymmetrical westward, gravity high of nearly 50 milligals over the anticlinorium relative to a broad low over the Middlebury synclinorium and Lake Champlain lowland (Figure 3). There is also a slight gravity high over the Adirondacks. The absolute Bouguer anomaly over the Green Mountains is only about -15 milligals which is anomalously high for continental regions. These results are similar to those of Diment (1968) and Bean (1953). It would appear that the structural relief displayed on the Precambrian unconformity from the floor of the synclinorium to the crest of the anticlinorium also affects the entire crust bringing dense material to high levels beneath the anticlinorium.

More detailed studies adjacent to the Champlain thrust by Smith (1972) show a distinct south-trending trough-like gravity low of about 8 to 15 milligals just east of the trace of the Champlain thrust east of Snake Mountain. The trough appears to turn southwesterly at the south end of Snake Mountain migrating to a similar position just east of the Orwell thrust. It thus crosses the southeasterly projection of the last outcrops of the Champlain thrust at nearly 90°.

Davidson(1970) carried out gravity studies and geologic mapping in the offset of the Champlain thrust between Buck and Snake Mountains (Stop 4) where an east-west cross fault was mapped by Welby (1961) and shown on the state geologic map (Doll and others, 1961). Welby estimated the down to the south displacement on this cross fault at about 2,000 feet to explain the westward offset of the Champlain thrust from Buck to Snake Mountain. Considering the trace of the Orwell thrust less than 100 feet of dip separation on such a fault can be generated. Gravity contours trend northerly west of the thrust traces between Snake Mountain and Addison and northwesterly east of the thrust traces between Buck and Snake Mountain. These trends, combined with results from numerous com-

puter-generated gravity models, and the structural data, suggest the cross-fault is unnecessary.

REGIONAL ASPECTS

If the dip separation on the Champlain thrust belt is 10,000 to 20,000 feet or more it becomes difficult to get rid of this displacement without driving the master sole thrust of the system into and beneath the Green Mountain anticlinorium offsetting the Precambrian basement (Figure 3) (Powell, 1969; Tennyson, 1970). Judging from structure sections drawn across northern Vermont (Doll and others, 1961) and at the latitude of Brandon (Crosby, 1963) south of Middlebury others reach similar conclusions. Several efforts to determine if the Champlain thrust system re-appears out of the eastern flank of the synclinorium, passing in the air over the anticlinorium, have resulted in no evidence to support such an option. This suggests the Champlain thrust belt, the Middlebury synclinorium, and the Green Mountain anticlinorium are inescapably linked in a strongly west-vergent tectonic system which drove the western cratonic foreland beneath the Champlain thrust belt and the west-vergent synclinorium-anticlinorium couple. Silurian-Devonian rocks seem to be involved in some way east of the anticlinorium. Thus, the massive eastward underthrusting and resultant couple, which based on gravity data must have involved the entire crust, are presumably in part at least "Acadian".

At James Pasture (Stop 2) and elsewhere in the Middlebury synclinorium (Crosby, 1963; Soule, 1967) there is ample evidence of a northwest-trending early "ductile" deformation which produced recumbent flow-folds with penetrative axial surface "mineral" cleavage. These structures are clearly deformed by later north-trending folds and cleavage (Stop 2), generally of a more "brittle" aspect. A similar fabric is found in schists on the Green Mountain anticlinorium (Tennyson, 1970). The later deformation appears to produce some, but not all, of the main map pattern of the synclinorium and anticlinorium. If the axial surface cleavage in folds found below the Champlain thrust is the same as the "late" cleavage it often resembles in the synclinorium (which has not been proved), it would appear the thrust belt developed sequentially west to east during and after most of the folding on both sides of the belt. Most workers conclude the early deformation was Taconic and relate it to emplacement of the Taconic "klippe". The later deformation is related to the Acadian and linked to the Champlain thrust and much of the gross geometry of the synclinorium-anticlinorium couple. Nothing in studies made thus far at Middlebury College refutes this general interpretation, but on the other hand neither does it prove it. Much remains to be done.

Finally, a most interesting problem is what happens to about 4,000 feet of Cambrian stratigraphy exposed in the Middlebury synclinorium which is missing in the Lake Champlain lowland. The

thrust belt appears to separate the two stratigraphies, and the thrust belt itself may have been controlled by the westward on-lap and thinning indicated by the facies change. Perhaps the initial thrust rode up a basement step on the base of one or another of the massive quartzite or dolostone struts breaking upward and out as the layers end to the west. On-going and future work will hopefully clarify this and many other interesting problems.

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ROAD LOG B-4

Note: Exercise caution while driving on narrow, dirt, country roads, and at blind, busy intersections. Parking space at Stops is generally limited. Park off road and do not block farm entrances or gates. Road Log starts from parking lot east of Middlebury College Science Center.

<u>Miles</u>

0.0 Leave Science Center parking area heading north, turn left onto Franklin Street which turns quickly north toward Route 125.

0.1 Stop. Turn right onto Route 125.

0.2 Turn left onto Weybridge Street.

1.0 Turn right following signs to Morgan Horse Farm. Chipman Hill at 1 o'clock. Chipman Hill rises 400 feet above the surrounding terrain and exposes no bedrock anywhere on its slopes. Gravity work over the feature suggests it is entirely of unconsolidated material, thus a glacial deposit, presumably a kame or drumlin.

1.4 Bear left. Road to right crosses Pulp Mill covered bridge.

1.8 Otter Creek on right.

2.7 Morgan Horse Farm on right.

3.3 Road intersection. Continue straight.

4.4 Descend into valley of Otter Creek.

4.6 On skyline at 3 o'clock, axis of Green Mountain anticlinorium.

4.9 Cross iron bridge over Otter Creek at Weybridge-New Haven town line. Power dam on left at Huntington Falls.

5.3 Stop 1: Folds in Middlebury synclinorium. Park along right side of road near power pole. Cross fence and descend across small ravine. Proceed up low rise to east to pasture and low hill about 1,000 feet east of road. Exposed on the hill, and in adjacent woods is a series of north-trending westward inclined folds with a well developed north-trending axial surface cleavage in limestones and dolostones of the Chipman Formation. Several prominent dolostone layers and the distinctive banded Weybridge Member enable the fold geometry to be clearly seen. These structures are interpreted by Crosby (1963) as late structures related to the gross geometry of the Middlebury synclinorium and probably Acadian in age.

5.3 Return to car, continue north up hill.

5.4 Turn sharp left at intersection.

6.0 Snake Mountain ridge at 12 o'clock on skyline.

6.2 Descend into valley of Otter Creek. Profile of Buck Mountain at 3 o'clock on skyline.

6.9 Turn sharp left at road intersection.

7.2 Turn left at intersection. Proceed across twin-bridges below Weybridge power dam and bear right. Continue south through Weybridge Village.

9.1 Weybridge Elementary School on left; cemetery on right.

9.3 Stop at yield sign. Then bear left onto Route 23 for 50 yards, turn left onto gravel road just past church on left.

9.9 Stop 2: Two deformations. Park at road intersection just before large white barn with two silos. Enter gate to pasture on northwest side of intersection. Please close gate after entering pasture. Proceed northwest across pasture toward low wooded ridge. No hammers, please.

This pasture, which lies in the core of the Middlebury synclinorium, was mapped by plane table by Crosby (1963) and has since been mapped hundreds of times by students from Middlebury College and numerous other institutions in New England. It is certainly one of the finest displays of multiple deformation geometry and fabric available for instructional use, and is probably one of the key areas for interpretation of the tectonic evolution of the Middlebury synclinorium.

Exposed in the pasture are a number of very subtle northwest-trending recumbent folds with a distinct axial surface mineral cleavage. The geometry of these folds is outlined by the contact between the base of the Middlebury Formation grey marbles and the top of the Beldens Formation dolostones and marbles. A thin grey slate just above the base of the Middlebury is a most useful marker. These folds and related cleavage are clearly deformed by a north-trending set of small folds and a prominent axial surface "crenulation" cleavage. The assumption is that the late folds in this pasture are the same generation as the folds seen at stop 1. The early folds here are interpreted as "Taconic".

9.9 Return to cars, turn around and retrace route to Route 23.

10.4 Stop. Turn right onto Route 23, bear to left of cemetery.

10.8 At 9 o'clock, valley of Lemon Fair River. Beyond wide valley is the dip slope on Monkton Formation off the east side of Snake Mountain. The Champlain thrust lies just on other side of skyline ridge dipping eastward beneath the Middlebury synclinorium.

11.1 Profile of Buck Mountain at 12 o'clock on skyline.

11.5 Slow for sharp turn to left in road.

13.4 Bridge over Lemon Fair River at confluence with Otter Creek on right.

14.5 First of many outcrops of Monkton Formation red quartzites and shales at base of dip slope of Snake Mountain.

15.1 Stop. Turn right onto Route 17.

15.4 Bridge across Otter Creek. Prepare for dangerous left turn on blind hill dead ahead.

15.6 Slow by white barn on right, then turn left off Route 17 onto road leading north, just before red brick farmhouse on northeast side of intersection.

16.7 Buck Mountain at 12 o'clock. The Champlain thrust lies just below the cliffs west of summit placing Monkton quartzites over Ordovician limestones and shales. The gentle east slope is a dip slope on Monkton Formation.

18.3 Stop 3: Champlain Thrust. Park at cemetery. Walk east along north side of cemetery fence to east end of cemetery. Cross fence and enter juniper woodland and pasture and proceed northeastward toward summit of Buck Mountain. The best route to the summit is up the northwest ridge on skyline through grove of large hemlock trees. The walk to the summit first crosses outcrops of Orwell limestone on the west flank of an overturned syncline. Glens Falls Formation, largely covered, is in the core. A minor thrust fault places Middlebury Formation limestones over the east flank of the fold. East of the thrust the Middlebury limestones are caught in a tight overturned anticline with a well developed axial surface fracture cleavage. An overturned syncline follows to the east with a core of Glens Falls Formation. The axis of the fold is just above the base of the steep slope rising to the summit at about the 550-foot contour line. Continuing up the steep slope outcrops of Orwell appear followed quickly by Middlebury limestone, all here overturned to the west. Directly above the Middlebury limestone are outcrops of Monkton quartzites and dolostones, followed above and to the summit by red Monkton quartzites. The Champlain thrust lies between overturned Middlebury limestone and right side up Monkton quartzite. The trace of the thrust is exposed to the south at the foot of the quartzite cliffs. Excellent panoramic view from the summit: Snake Mountain to the southwest, the Lake Champlain Lowland and the Adirondack Mountains to the west. Retrace route to cemetery and cars.

18.3 Turn around and return to Route 17.

18.4 Profile of Snake Mountain at 1 o'clock. Champlain thrust lies at base of steep cliffs and slopes just below skyline and dips about 7° east.

20.9 Slow for right turn onto Route 17.

21.3 Continue straight on Route 17 past junction from left of Route 23.

21.5 Stop 4: Orwell Thrust. Park along north side of road near gate into pasture on right. Walk west along road to outcrops of Bridport dolostones on right. Just beyond to west is a large roadcut in Ordovician Stony Point shale. The Orwell thrust lies between the two exposures placing Bridport over Stony Point. If one enters woods and as-

cends hill north of road the fault trace can be followed to the summit of the hill in a pasture where Bridport caps the west face of the hill over shale. Good exposures of Bridport Formation in pasture. Looking south-southwest across road and valley to Snake Mountain from summit of hill the Orwell thrust lies several feet into woods at edge of hillside pasture. The Champlain thrust lies just above it descending from summit of Snake Mountain far to west. The Champlain thrust truncates the Orwell thrust several hundred yards to west and Monkton quartzite lies directly on shale from here to the west face of Snake Mountain. Note similar elevation of thrust traces north and south of road suggesting minor, if any, cross faulting. Return to cars.

21.5 Continue west on Route 17.

23.0 At 9 o'clock the Champlain thrust is exposed at base of quartzite cliffs near skyline ridge. Several caves which have formed in weak shale below the resistant quartzite provide excellent exposures of the overhanging fault plane and minor structures in shale below.

23.3 Turn left off Route 17 onto gravel road leading south along west face of Snake Mountain.

26.6 Turn left (east) on gravel road.

27.1 At 3 o'clock, Crane School hill half a mile to south is either a klippe or tongue-like salient of the Orwell thrust placing Bridport and Crown Point Formations over Ordovician shales. At 10 o'clock, just to left of saddle on Snake Mountain ridge on skyline the Orwell and St. George thrusts are truncated by the Champlain thrust.

28.0 Turn right at intersection. Orwell thrust "klippe" straight ahead.

28.5 At 3 o'clock cliffs of folded Bridport Formation in woods just above Orwell thrust.

28.8 Beehives on left. On right, lower slopes in pasture expose Ordovician shales below Orwell thrust.

29.0 Slow for road junction. Turn left.

29.2 Iberville shale exposed on knolls and hills surrounding farms.

29.6 At about 10 o'clock Champlain thrust lies at foot of prominent cliffs of Monkton quartzite at edge of farmland.

29.7 Culvert across small creek. From this point to pass over Snake Mountain at 30.2 the route crosses all three thrusts.

29.9 Road crosses Champlain thrust. Monkton Formation is exposed in small road cuts and in quarry at right on upper plate.

30.2 Pass over Snake Mountain. Shift to lower gear for steep, winding descent.

30.6 Crossroads. Turn right (south).

30.8 Winooski Formation exposed in cliffs on east side of road.

31.9 First of many dip slope outcrops of Monkton Formation.

32.4 Stop. Intersection with Route 125. Cemetery on right. Turn right. Around this intersection are the last and southernmost exposures of the Monkton Formation and the Champlain thrust which presumably angle southeasterly beneath glacial cover.

32.8 At 3 o'clock cliff of well-bedded Monkton quartzites. At 12 o'clock, and extending south, low hills are in Bridport and Crown Point Formations on upper plate of Orwell thrust here heading southwest.

33.3 Continue on blacktop past dirt road to right. From here to 34.0 road passes complex relationships involving Orwell and St. George thrusts. At 33.5 road cuts on right expose Glens Falls Formation on upper plate of St. George thrust which angles nearly east-west up the brow of hill to north. On left at 33.9 outcrops of Bridport Formation on upper plate of Orwell thrust.

34.5 Turn right (north) off Route 125 onto narrow blacktop.

35.0 Intersection. Turn right. Good panorama of Snake Mountain to northeast.

35.8 Stop 5: St. George Thrust. Park in larger quarry in Ordovician shales on right. Brief stop to examine the St. George thrust and minor structures in lower plate, and to view problem of the southern termination of the Champlain thrust. From quarry ascend into woods and then pasture to top of hill. Glens Falls Formation outcrops on north side of hill on upper plate of St. George thrust. From top of hill cliffs of Monkton Formation to east can be seen plunging southeastward into Lemon Fair valley. Champlain thrust is at foot of cliffs and hills. The thrust is lost from here southeastward beneath extensive glacial cover. Follow trace of St. George thrust east-northeast down hill to

stream and dirt road. Both Orwell and St. George thrusts continue north of road on west face of long south-plunging ridge in woods. Turn left on dirt road and return to cars.

35.8 Continue east along dirt road.

36.3 Stop. Blind intersection with busy Route 125. Turn left (east) on Route 125.

37.3 Slow down for sharp left turn on Route 125 after bridge over Lemon Fair River.

38.3 Straight ahead are The Ledges, a prominent north-trending cliff exposing Bascom through Middlebury carbonates on west flank of Middlebury synclinorium.

38.7 Slow for left bend and winding ascent of The Ledges.

39.8 Slow for right bend in road at ridge crest.

40.0 Slow for hidden crossroad. Hortonville shales in core of Middlebury synclinorium exposed in small roadcuts and outcrops.

41.8 Blinker light at edge of Middlebury College campus.

42.0 Turn right opposite Catholic Church onto Franklin Street and Middlebury College Science Center. End of trip.

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Trip B-5

ANALYSIS AND CHRONOLOGY OF STRUCTURES ALONG THE CHAMPLAIN THRUST WEST OF THE HINESBURG SYNCLINORIUM

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INTRODUCTION

The Champlain thrust has long attracted the attention of geologists. Prior to the discovery of fossils along this belt the thrust was considered an unconformity between the strongly-tilted Ordovician shales of the "Hudson River Group" and the overlying, gently-inclined dolostones and sandstones of the "Red Sandrock Formation" (Dunham, Monkton, Winooski formations of Cady, 1945). The "Red Sandrock Formation" was thought to be Silurian in age since it was lithologically similar to the Medina Sandstone of New York. Between 1847 and 1861 fossils of pre-Medina age were found in the "Red Sandrock Formation" and its equivalent "Quebec Group" in Canada. Based on this information, Hitchcock (1861, p. 340) stated that "it will be necessary to suppose the existence of a great fault, extending from Quebec through the whole of Canada and Vermont and perhaps to Alabama. Its course through Vermont would correspond very nearly to the western boundary of the red sandrock formation." Since then, although its extent has been greatly limited, its importance has not diminished.

Our understanding of the Champlain and associated thrusts is primarily the result of studies by Keith (1923, 1932), Clark (1934), Cady (1945), and Welby (1961). Current interest in earthquake research on the character, movement, chronology, and mechanics of faults requires a closer, more detailed study of such well mapped faults as the Champlain thrust.

Acknowledgements

The work of Cady (1945) and Welby (1961) along the Champlain thrust in western Vermont is very valuable in providing the framework for detailed work that is presently being done at the University of Vermont and Middlebury College. Although many geologists have contributed to our present understanding of this region, syntheses by Cady 1969, Doll and others 1961, Rodgers 1968, and Zen 1967, 1968 are particularly helpful.

Many students at the University of Vermont have contributed information for the localities in this trip. Data on fractures, faults, and quartz deformation lamellae at the Shel-

burne Access Area were collected by Charles Rubins, John Millett, Edward Kodl, Robert Kasvinsky, Evan Englund, and Jack Chase. The analyses at locality S9 and Mount Philo are largely the work of Arthur Sarkisian. Richard Gillespie, Roger Thompson, Jack Chase, Greg McHone, and Gary French provided information for Pease Mountain.

REGIONAL GEOLOGY

The Champlain thrust extends for approximately 75 miles from Cornwall, Vermont, to Rosenberg, Canada, and places Lower Cambrian dolostone with some quartzite on highly deformed Middle Ordovician shale and minor beds of carbonates. Throughout its northern part the thrust is confined to the lower member (Connors facies) of the Dunham Dolostone. At Burlington the thrust apparently rises 2000 feet in the section to the dolostones and quartzites of the lower part of the Monkton Quartzite. It then truncates major structure in the Ordovician rocks of the lower plate south of Mount Philo to Cornwall, Vermont (Doll and others, 1961). The upper part of the Dunham only reappears along the thrust just south of Buck Mountain (Welby, 1961). The Champlain thrust appears to be primarily restricted to the first massive dolostone interval above the Precambrian.

Throughout most of its extent north of Pease Mountain (figure 1) the trace of the Champlain thrust is somewhat straight and the surface strikes to the north and dips gently to the east at angles less than 20 degrees. South of Pease Mountain the trace of the thrust is irregular because of subsequent folding and faulting (Doll and others 1961, Welby 1961). At Mount Philo, for example, the rocks of the upper plate are folded gently into an east-plunging syncline. Between Burlington and Snake Mountain the thrust is cut by a number of cross-faults that are interpreted as normal faults by Welby (1961, p. 204). Our work indicates that the displacement on some of these faults is predominantly strike slip (Stanley 1969, Sarkisian 1970).

The stratigraphic throw on the Champlain thrust is in the order of 8000 feet at Burlington. To the north the throw decreases as the Champlain thrust is lost in the shale terrain north of Rosenberg, Canada. Part, if not all, of this displacement is taken up by the Highgate Springs and Philipsburg thrust which continues northward and becomes the "Logan's Line Thrust" (Cady 1969). South of Burlington the stratigraphic throw is in the order of 5500 feet. As the throw decreases on the Champlain thrust near Cornwall the displacement is again taken up in part by the Orwell, Shoreham, and Pinnacle thrusts, which place Upper Cambrian and Lower Ordovician rocks on each other and eventually on the Middle Ordovician rocks to the west (Cady 1945).

The configuration of the Champlain thrust at depth is speculation. Where it is exposed the thrust surface is essentially parallel to the gently-dipping beds in the upper plate. This thrust geometry persists for at least 2-3 miles east of its most western limit since the thrust is still essentially parallel to the bedding in the Monkton Quartzite at the base of the upper plate in the center of the recess of the thrust trace on the Monkton culmination. Further to the east the thrust must eventually steepen in dip and pass into Precambrian basement since it does not reappear at the surface on the west side of the Precambrian core of the Green Mountains. This overall configuration is shown in the cross sections accompanying the geologic map of Vermont (Doll and others 1961).

The age of the Champlain thrust is debatable. Cady (1969, p. 75) believes the thrust developed in the Acadian Orogeny although the youngest rocks exposed below the Champlain thrust are Middle Ordovician in age. Welby (1961, p. 221) believes the thrust developed during the Taconic orogeny of Middle to Upper Ordovician age. Thrusting predates the emplacement of the Mesozoic dikes which clearly cut the structures of the Champlain thrust.

Our work shows that the Champlain thrust has undergone an extensive structural history involving possibly more than one period of thrusting. Perhaps the most compelling evidence for a multiple history of displacement is the presence of prograde chlorite in recrystallized fractures in the Monkton on the upper plate and the absence of prograde chlorite throughout the pelitic rocks directly below the thrust. We tentatively suggest that the Champlain thrust was originally developed during the Taconic orogeny, metamorphosed and then reactivated to place a metamorphosed upper plate on an unmetamorphosed lower plate. Although the second event may also be Taconic in age, since rocks of Silurian and Devonian age are not present below the thrust, the additional structural events may have occurred during the Acadian orogeny. These speculations will be discussed during the trip.

STRATIGRAPHY

A composite stratigraphic section and correlation chart for the area of the Champlain thrust and the Hinesburg synclinorium are shown in Table 1 and Table 2 respectively.

TABLE 1.

COMPOSITE STRATIGRAPHIC SECTION FOR THE HINESBURG SYNCLINORIUM AND AREA

WEST OF THE CHAMPLAIN THRUST AND EAST OF THE HINESBURG THRUST

(Cady, 1945; Doll, et al., 1961)

Middle Ordovician

Hathaway Formation*

Missing but present north of the Lamoille River.

Iberville Formation

Noncalcareous shale, rhymthically interbedded with thin beds of silty dolomite and in the lower part with calcareous shale.

Stony Point Formation

Calcareous shale that grades upward into argillaceous limestone and rare beds of dolomite.

Cumberland Head Formation

Missing but well exposed in Grand Isle County.

Glens Falls Formation

Thin bedded, dark blue-gray, rather coarsely granular and highly fossiliferous limestone.

Orwell Limestone

Missing but present south of Charlotte.

Middlebury Limestone

Missing but present south of Charlotte.

Lower Ordovician

Chipman Formation

Missing in the Hinesburg synclinorium but present south of Charlotte.

Bascom Formation

Beds of light bluish gray calcitic marble with laminae and thin beds of siliceous phyllite. Beds of brown-orange weathering dolomite 1 to 3 feet thick. The upper part becomes more phyllitic and is mapped separately as the Brownell Mountain Phyllite. Contact against typical Cutting dolomite is gradational.

Massive whitish to light grayish weathering dolostone, dark gray on the fresh surface. Sand-size quartz grains present in places especially near the base where sandy laminae are more abundant and brecciated in places. Black chert nodules are found in the upper part. Contact is sharp with sandy dolomite above and white calcitic marble of the Shelburne Formation below.

Shelburne Formation

Massive whitish gray weathering calcitic marble that is white on the fresh surface. Laminae of green phyllite present in the eastern part of the Hinesburg synclinorium. Contact is sharp with typical calcitic marble above and gray dolomite with quartz knots below.

Upper Cambrian

Clarendon Springs Dolomite

Massive, gray weathering dolomite with numerous knots of white quartz crystals. Black chert commonly found in the upper part. Contact gradational with the Danby Formation.

Danby Formation

Beds of gray sandstone interlayered with beds of dolomite 1 to 2 feet thick and sandy dolomite 10 to 12 feet thick. Sandstones are cross laminated. Beds 1 to 3 feet thick. Contact gradational.

Cambrian

Underhill Formation

Fairfield Pond Member: Predominantly green quartz - chlorite - sericite phyllite. Quartz grains common.

White Brook Member: Chiefly brown-weathered whitish, tan and gray sandy dolomite.

Pinnacle Formation

Schistose graywacke, gray to buff, with subordinate, quartz-albite-sericite-biotite-chlorite phyllite. Includes the Tibbit Hill Volcanic Member.

*Only those formations encountered in the course of the trip will be described. Kindly refer to the Centennial Geologic Map of Vermont for other formation descriptions.

System	Series	Group	Saratoga Springs, New York Mohawk Valley (Fisher 1965)	West-central Vermont (Doll et al 1961)	West Limb of the St Albans Synclinorium (Doll et al 1961) (Shaw 1958)	Lincoln Mtn - Enosburg Falls Anticline East of Hinesburg Thrust (Doll et al 1961) (Stone and Dennis 1964)
ORDOVICIAN	Middle	Trenton	Schenectady Shale Canajoharie Shale Shoreham Limestone Larrabee Limestone	Hathaway Formation Iberville Formation Stony Point Formation Cumberland Head Formation Glens Falls Formation	Morse Line Formation	
		Black River	Amsterdam Limestone	Orwell Limestone		
		Chazy		Middlebury Limestone		
	Lower	Beekmantown	Chuctanunda Creek Dolostone Gailor Dolostone	Chipman Formation Bascom Formation Cutting Dolomite Shelburne Formation	Highgate Formation	
CAMBRIAN	Upper		Hoyt Limestone Mosherville Sandstone Theresa Formation Potsdam Sandstone	Clarendon Springs Dolomite Danby Formation	Gorge Formation Sweetsburg Fm.: Hungerford Sl., Rockledge Congl., Sweets Corner Sl., Mill River Congl., St. Albans Sl.	
	Middle					
	Lower			Winooski Dolomite Monkton Quartzite Dunham Dolomite Cheshire Quartzite	Rugg Brook Formation Parker Slate Dunham Dolomite Champlain Thrust	Dunham Dolomite Cheshire Quartzite
CAMBRIAN?				Mendon Formation	Not Exposed	Underhill Formation Pinnacle Formation
PRECAMBRIAN			Metamorphic Rocks of the Adirondack Dome	Mount Holly Complex		Mount Holly Complex

Table II Correlation chart for selected areas in western Vermont and eastern New York.

STOP DESCRIPTIONS

General

The trip consists of five stops along the Champlain thrust. These stops are located on figure 1.

Stop 1. Lone Rock Point, Burlington (1, 2a, 2b, figure 2) - This locality is perhaps one of the finest exposures of the Champlain thrust in Vermont and Canada. Here the Dunham Dolomite (Conners facies) of Lower Cambrian age overlies the Iberville Formation of Middle Ordovician age. The thrust contact is sharp and marked by a thin zone of breccia in which angular clasts of dolostone are embedded in a highly contorted matrix of shale. Slivers, several feet thick, of limestone are found along the fault and may represent pieces of the Beekmantown Group (Beldens Member of the Chipman Formation ?). The undersurface of the Dunham Dolomite along the thrust is grooved by fault mullions which plunge 15° to the southeast (figure 2, diagram 1 and 2a). The average southeastward dip of the thrust is 10 degrees.

A variety of minor structures are found in the Iberville Formation whereas fractures are the only structures in the Dunham Dolomite. Many of these faults are filled with calcite and grooved with slickensides. The minor folds in the shale are numerous, and are easily grouped into two ages. The early folds have a well developed slaty cleavage that forms the dominant layering in the Iberville and is concordant to the thrust surface at the base of the Dunham Dolomite.

The younger generation consists of asymmetrical drag folds with short, gently curved hinges and rather open concentric profiles. The axial surface is rarely marked by cleavage but when it is developed, fracture cleavage, filled with calcite, is typical. These folds deform the slaty cleavage of the older generation and are related to movement of the Champlain thrust since they decrease in abundance away from the thrust surface. The orientation of 59 axes with their sense of rotation is shown in diagrams 1, 2a, and 2b of figure 2.

Slip line orientations. Drag folds of the younger generation have been used at 5 localities to determine a direction of movement on the Champlain thrust. Two of these localities are in the Iberville or Stony Point Formations directly below the thrust and 2 localities are in the Monkton Quartzite just above the thrust (diagrams 1, 2a, 2b, 3, 7, 8, figure 2). The remaining locality at Shelburne Point is along a fault zone in the Stony Point Formation less than a mile west of the Champlain thrust. At each of these localities numerous younger folds are developed with nonparallel hinges and short limbs facing in a variety of directions.

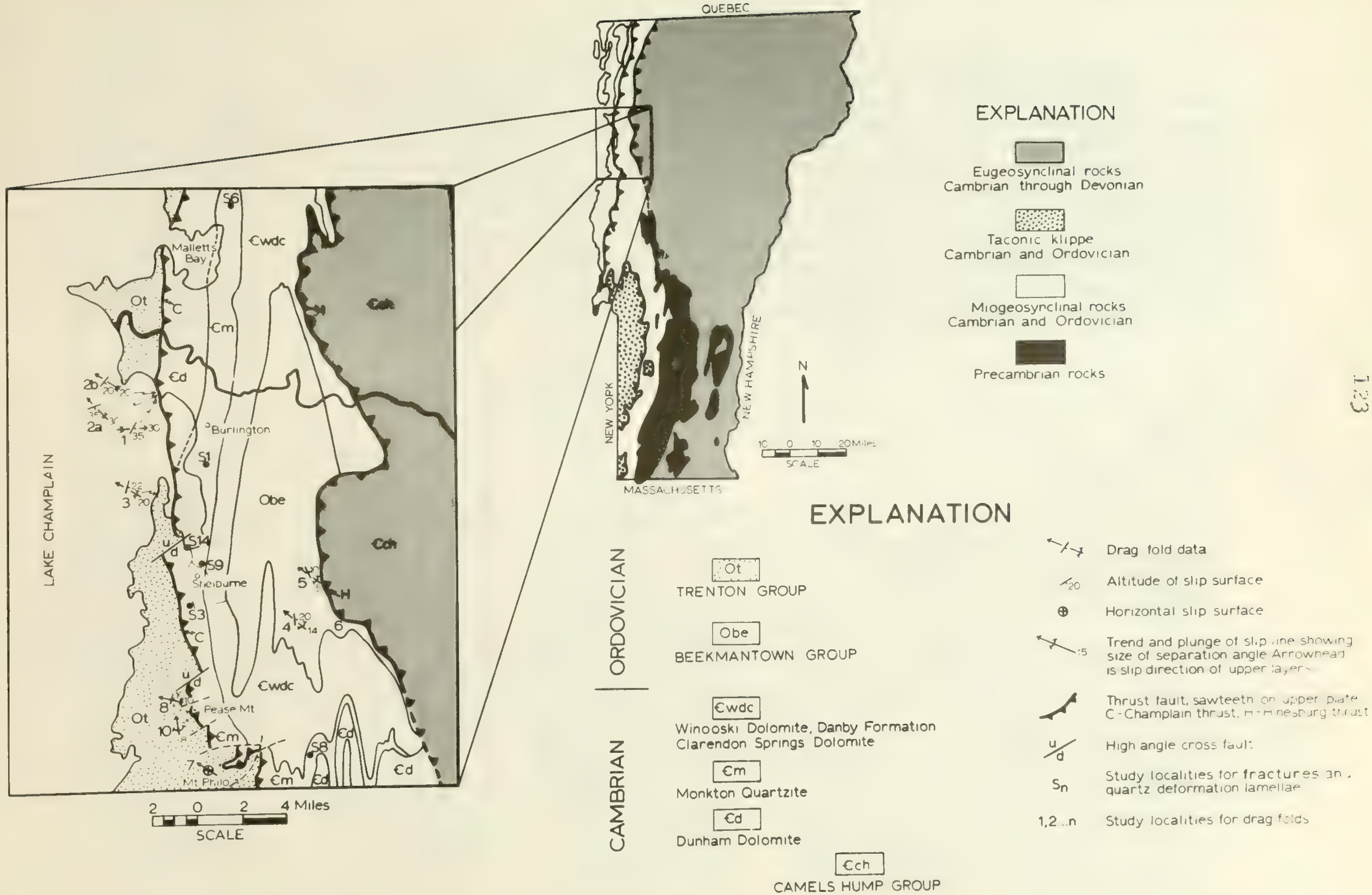

Figure 1. Generalized geologic map of the Hinesburg synclinorium. Slip line directions at localities 1-8, and 10 are shown in more detail in figure 2.

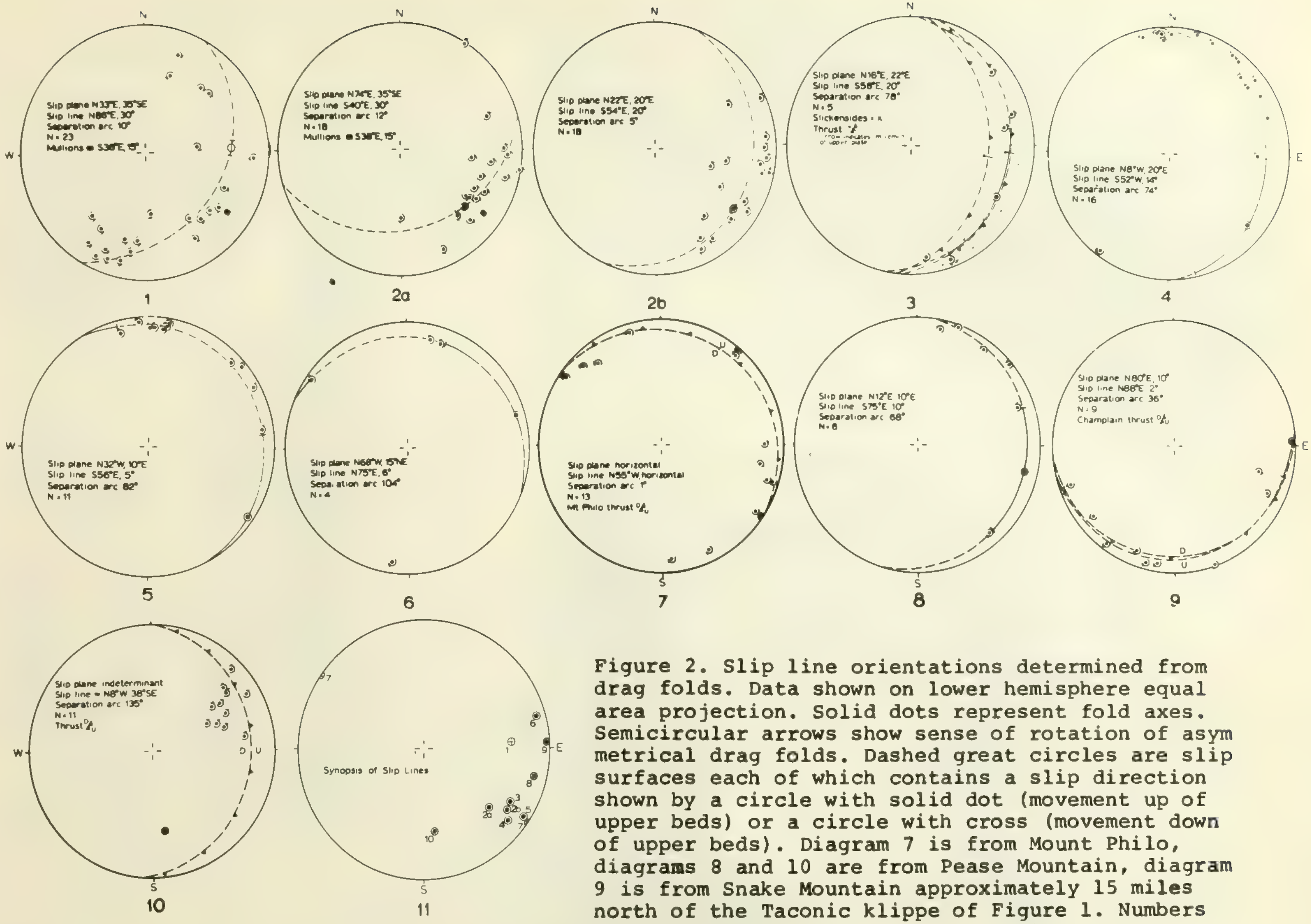

Figure 2. Slip line orientations determined from drag folds. Data shown on lower hemisphere equal area projection. Solid dots represent fold axes. Semicircular arrows show sense of rotation of asymmetrical drag folds. Dashed great circles are slip surfaces each of which contains a slip direction shown by a circle with solid dot (movement up of upper beds) or a circle with cross (movement down of upper beds). Diagram 7 is from Mount Philo, diagrams 8 and 10 are from Pease Mountain, diagram 9 is from Snake Mountain approximately 15 miles north of the Taconic klippe of Figure 1. Numbers correspond to locations shown on Figure 1.

A slip line or movement direction was determined at three stations along the 2000 feet of exposure at Rock Point (diagram 1, 2a, 2b, figure 2) using the methods described by Hansen (1967, 1971). The hinge orientation and sense of rotation for 18 to 23 younger folds were plotted at each station on a lower hemisphere equal area net. The great circle that best approximates the spatial distribution of axes defines the slip plane which is approximately parallel to the older cleavage and the Champlain thrust. At Rock Point this cleavage is of compact shale separated by thinner layers of extremely fissile shale. In all the diagrams in figure 2 clockwise or dextral folds occupy one part of the great circle, whereas counterclockwise or sinistral folds occupy the other. The arc that separates the opposite senses of rotation contains the slip direction. This is uniquely defined when the separation arc is zero. In most localities the separation arc is greater than zero and the bisector of the separation arc is arbitrarily designated as the slip direction. The overall symmetry of the fabric is monoclinic with the plane of symmetry oriented parallel to the slip direction and perpendicular to the slip plane.

The location of clockwise and counterclockwise arrows on either side of the separation arc indicates the direction of movement of the upper layers along the deduced slip line. In diagrams 2a and 2b (figure 2) the upper layers moved to the northwest approximately along a line striking N40W for 2a and N54W for 2b. In contrast , the upper layers moved eastward along a line striking N86E for tne southern part of the Champlain thrust at Rock Point (diagram 1, figure 2). In all three diagrams the separation arc ranges in size from 5 to 12 degrees.

Discussion of Results. The kinematic basis for drag fold analysis has been worked out in such geologic environments as tundra and sod slides, glacial lake clays, lava flows, and metamorphic rocks of all grades (Hansen 1967, Hansen and others 1967, Howard 1968, Hansen 1971). Scott (1969, p. 251-254) has verified these methods in the laboratory using substances of different viscosities. In all these studies it has been shown that the separation arc contains the slip line and that the drag folds are a product of one overall movement regimen.

As shown in diagram 11 of figure 2, the deduced slip lines are nearly parallel along 25 miles of the Champlain thrust. These slip directions are essentially parallel to fault mullions on the thrust surface at Lone Rock Point and slickensides on calcite-veneered surfaces at Shelburne Point (diagram 3) and elsewhere in the Middle Ordovician shale of the lower plate (Hawley 1957, p. 81). Although the origin of the diversity in hinge orientation in the rocks along the Champlain thrust is still unclear, the approximate parallelism among slickensides, mullions, and

slip lines indicates that the slip lines deduced from drag folds are a reliable movement indicator.

A generalized principal plane of stress and strain can be determined from slip line and plane information if rotation about the pole to the slip plane is assumed to be zero. With this restriction the slip line and the pole to the slip plane define the plane of σ_1 and σ_3 and λ_1 and λ_3.[1] This plane, known as the deformation plane, is also the plane of monoclinic symmetry of the drag fold diagrams. The location of σ_1 in the deformation plane depends on the sense of shear across the slip surface, the coefficient of internal friction of dolostone on shale and the strong planar anisotropy along the Champlain thrust.

The anomalous easterly slip direction for the southern part of Lone Rock Point (diagram 1, figure 2) will be discussed during the trip.

Stop 2. Shelburne Access Area (S14, figure 1) - The fractures and faults at this locality are ideal for dynamic analysis. The outcrop is located in the upper member of the Monkton Quartzite approximately 900 feet above the Champlain thrust. A high angle cross fault offsets the thrust just to the northwest (figure 1).

The location, orientation and relative displacement on faults and feather fractures are shown on the geologic map (figure 3). At each numbered station the orientations, relative abundance and surface features of 10 fractures were measured. Diagram A of figure 4 shows the poles to 248 fractures and diagram B shows four planes corresponding to the maxima in diagram A. The trend and deduced sense of displacement of the feather fractures are shown in diagram D of figure 5.

The faults in figure 3 are generally vertical, contain a very narrow zone of gouge and are divided into an east-west group and a north-south group according to their general strike. The north-south group are few in number and displace the east-west faults and hence, are younger in age. The faults of both generations are wrench faults since the dip slip displacement is only several inches and the strike slip displacement is as large as 3 feet. Furthermore, feather fractures adjacent to many of the faults are only present on the horizontal surfaces. Petrofabric analysis of quartz deformation lamellae in samples M1, M3, M4 (figure 4) further supports this conclusion (figure 6).

[1] σ_1, σ_2 , σ_3 refer to the principal axes of stress with σ_1 representing the maximum compressive stress. λ_1, λ_2 , λ_3 are the principal directions of quadratic elongation with λ_1 representing the direction of maximum elongation.

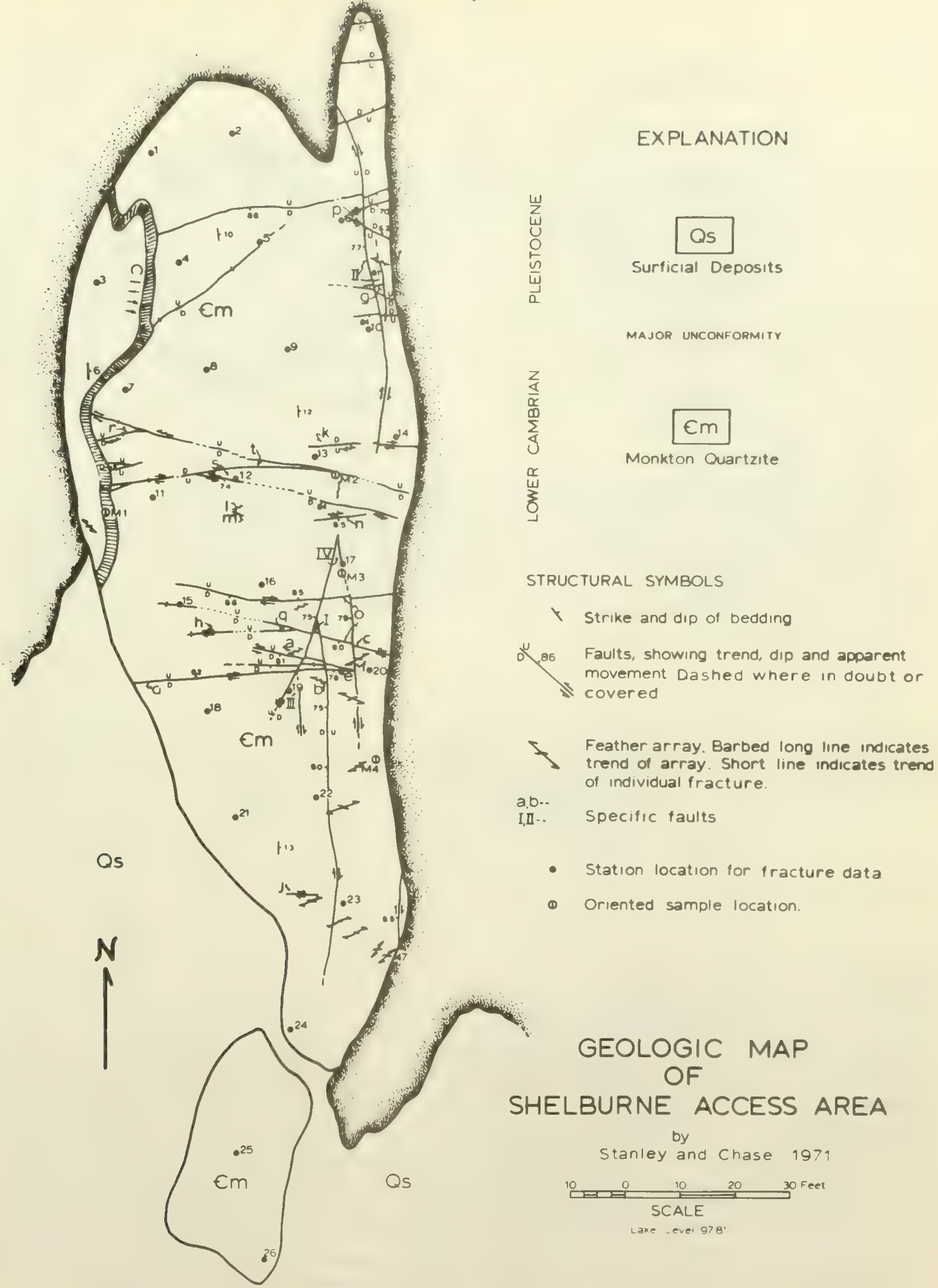

Figure 3. Structures in the Monkton Quartzite at the southern end of Shelburne Bay, western Vermont. Located at station S14 on figure 1.

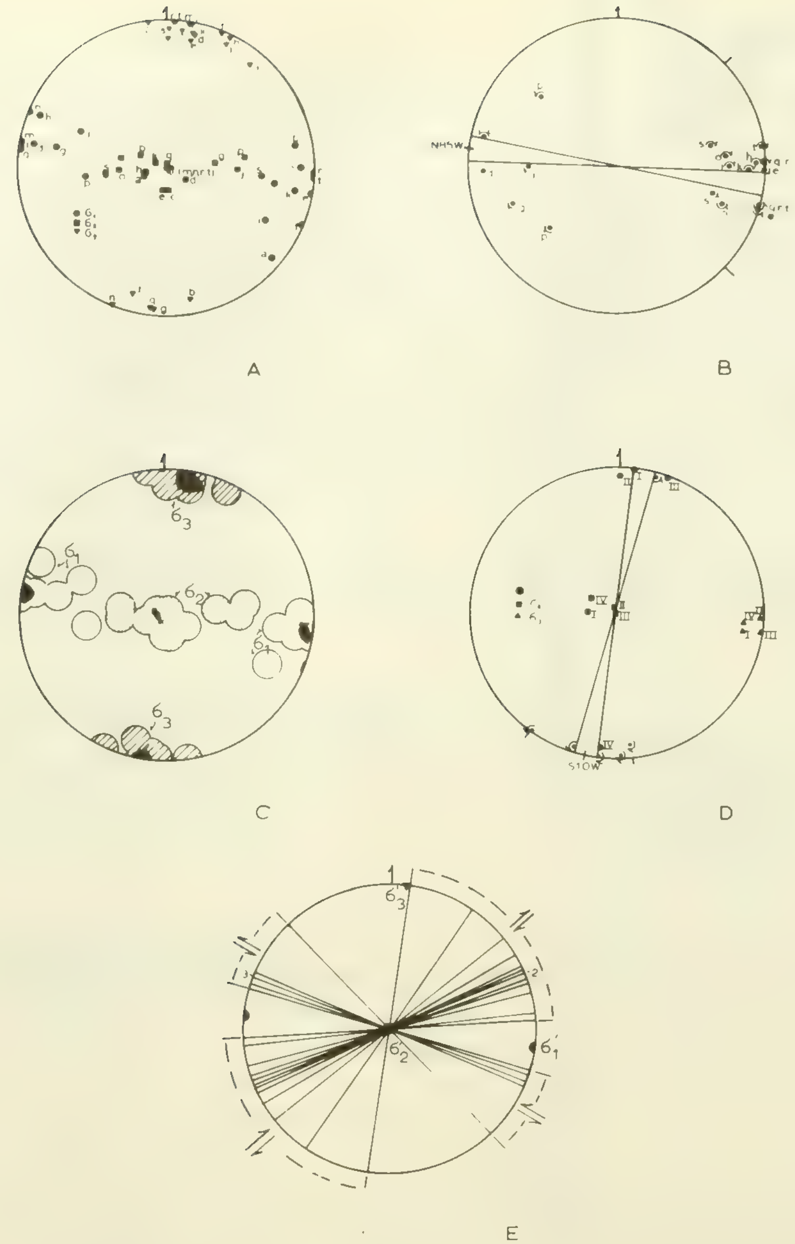

Figure 5. Lower hemisphere equal area diagrams of faults and feather fracture arrays at Shelburne Access area. Diagram A shows the principal stresses deduced for the complimentary faults (e,o,p,q) and fault-feather fracture sets (c,d,g,h,i,j, k) of the first generation of wrench faults in figure 3. These directions are contoured in diagram C. Contour interval 9.1, 18.2 percent per 1 percent area. Diagram B shows the slip directions (solid dot) with their respective senses of shear for each of the faults in diagram A. Diagram D shows the principal stresses deduced for the second generation of complimentary wrench faults. Their respective slip directions and senses of shear are indicated by a solid dot with concentric arrows. Diagram E shows the trend and relative displacement deduced from 21 feather-fracture arrays.

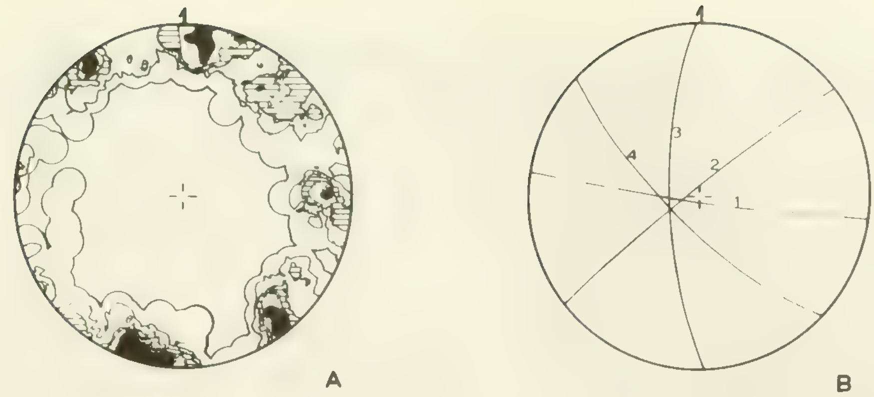

Figure 4. Lower hemisphere equal area projections of macroscopic fractures in the Monkton Quartzite at the Shelburne Access area. Diagram A shows 248 poles to fractures. Contour intervals are 0.4, 1.2, 2.0, 2.8, 3.6 respectively per 1 percent area. Diagram B shows planes corresponding to the 3.6 percent maxima of diagram A.

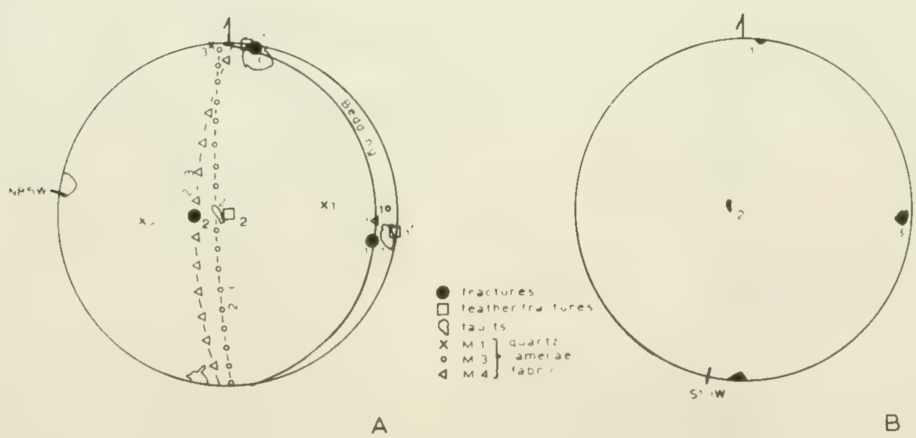

Figure 6. Synoptic diagram of the principal stress positions deduced for generation one and two wrench faults, megascopic fractures, feather fracture arrays and quartz deformation lamellae at Shelburne Access area.

The dihedral angle between complementary wrench faults (for example, e, p, i figure 3) ranges from 15 to 70 degrees with an average of 27 degrees which implies either a high value for the angle of internal friction, or fracturing under low effective confining pressures (less than 500 bars perhaps). These conditions were probably near the earth's surface since the pressure effect of a pore fluid was minimal after low grade metamorphism.

Microscopic planes of hematite inclusions, recrystallized quartz veins, unfilled fractures, quartz deformation lamellae, and undulose extinction in quartz pervade all thin sections. Prograde chlorite occurs between grains and along recrystallized quartz veins. Thus, the early sets of fractures were metamorphosed at the chlorite grade forming the recrystallized quartz veins and planes of inclusions. The unfilled fractures and quartz deformation lamellae were then superposed on and influenced by this annealed fabric.

Dynamic Analysis. Fractures, faults and quartz deformation provide information on the orientation and relative magnitudes of the principal stresses.

1) Fractures: Dynamic interpretation of fractures is based on geometry and the identification of shear or extension fractures. The intersection of the 4 fracture sets in diagram B (figure 4) defines the σ_2 position. The σ_1 direction is oriented 90 degrees to σ_2 in the plane that bisects the acute angle between shear fractures. The acute angles between fracture sets 1 and 3 and 2 and 4 are 80 and 83 degrees respectively. Set 1 fractures are in the extension position because they parallel the fractures in the feather arrays. Fracture sets 2 and 4 are, therefore, shear fractures and σ_1 plunges eastward at 10 degrees in the plane of fracture set 1. σ_3 trends northward and corresponds to the pole of fracture set 1. The deduced principal stresses are compatible with the stress configuration indicated by the wrench faults of generation 1.

2) Feather fractures: The feather arrays in diagram D (figure 5) with their respective senses of shear indicate that σ_1 is oriented east-west, σ_3 trends north-south and σ_2 is vertical. The principal stresses deduced for fractures and feather arrays agree in trend but differ by 10 degrees in dip since only the trend can be measured and not the dip of feather arrays.

3) Wrench faults of generation 1: Diagram A in figure 5 shows the deduced positions for the principal stresses calculated for faults labelled a through q on figure 3. These calculations were based on complementary faults (e,o,p,q) and faults with their associated feather fractures (c,d,g,h,i,j,k). The principal

stress positions are contoured in diagram C of figure 5 which shows that σ_1 plunges gently eastward, σ_3 trends northward, and σ_2 is nearly vertical. Diagram B (figure 5) shows that right lateral faults trend northeasterly whereas left lateral faults trend northwesterly.

4) Wrench faults of generation 2: The deduced positions for the principal stresses calculated from complementary faults (I, III, and IV, figure 3) and offset structures cut by fault II (figure 3) are shown in diagram E of figure 5. A comparison of diagrams C and E of figure 5 indicates that σ_2 for fault generations 1 and 2 are parallel. The positions of σ_1 and σ_3 however are interchanged. This orthogonal relationship implies that the second generation may have been caused by displacements associated with the first generation. As movement occurred during generation 1 the east-west stresses were reduced to a minimum value and the north-south stresses were simultaneously increased to the maximum compressive value. The stage was then set for generation 2 faulting.

5) Deformation lamellae in quartz: The lamellae are similar in character to those described by Carter and others (1964). The results were analyzed using methods described by Carter and Friedman (1965), and Scott and others (1965). The deduced orientation for σ_1, σ_2 , and σ_3 are shown in figure 6. In M3 and M4, σ_1 lies in the bedding and σ_2 appears to be equal to σ_3 in magnitude. In Ml, σ_1 is inclined 40 degrees to the east, σ_2 dips 50 degrees to the west and σ_3 trends northward and is horizontal. Although these orientations are not parallel in all samples, they are consistent with the stress positions deduced from the fractures, feather fractures, and first generation faults. The stress configuration in Ml is triaxial with $\sigma_1>\sigma_2>\sigma_3$ whereas the configuration in samples M3 and M4 is biaxial with $\sigma_1>\sigma_2 = \sigma_3$. These patterns indicate that the quartz lamellae formed during and slightly after the wrench faults of generation 1.

Relationship to major faults. Wrench faults are commonly associated with thrust faults. Both can be related to the same σ_1 direction and only require a switch of σ_2 and σ_3 in the stress configuration during thrusting to develop wrench faults. The small wrench faults in the Monkton Quartzite bear the same relationship to the Champlain thrust and as such, suggest that some cross faults shown on the Geologic Map of Vermont (Doll, and others, 1961) are indeed wrench faults.

One of these cross faults was mapped by Welby (1961) just to the northwest of the Shelburne Access Area (figure 1). It strikes northeasterly and displaces the upper and lower plates of the Champlain thrust in a right lateral direction. Because the deduced sense of σ_1 for first generation faults and assoc-

iated structures dip eastward more gently than the Champlain thrust (approximately 10 degrees) the inferred horizontal displacement on the cross fault would result in the same apparent vertical movement as indicated on the map (figure 1). This movement geometry also characterizes the right lateral faults of generation 1 at Shelburne Access Area. The cross fault of the west side of Shelburne Bay and the first generation faults and their associated structures are considered to be coeval, and therefore, younger than the Champlain thrust since the major cross fault clearly cuts both the plates of the thrust.

Structural History. The structural history for this outcrop and nearby major faults is summarized in figure 7.

Stop 3. Location S9 - Route 7 a mile north of Shelburne (S9, figure 1) - As shown on figure 8 the Winooski Dolomite is downthrown against the Monkton Quartzite along a normal fault trending slightly north of east. Eight smaller normal faults of similar trend cut parts of the Winooski (one is in the Monkton). Three small faults trend east of north and may be related to a larger fault which offsets the fault between the Monkton and the Winooski. This fault was apparently excavated when Route 7 was constructed. Several of the northeasterly-trending faults have well defined gouge zones ranging in thickness from less than an inch to slightly more than a foot. The most obvious zone is along the fault between the two formations. Well developed slickensides indicate a dominant dip slip component for all displacements (figure 9). A second nearly horizontal slickenside is present on the fault directly north of station 19 in the Monkton Quartzite on the east side of Route 7.

Dynamic interpretation. The synoptic diagram in figure 9 shows the orientation of the faults and associated slickensides. These normal faults indicate a state of stress in which σ_3 would be horizontal and trend northwesterly, σ_2 would parallel the general strike of the faults, and σ_1 would plunge to the southwest almost vertically. Inasmuch as the north-northeasterly faults cut the east-northeasterly faults the principal axes of stress probably rotated counterclockwise during this second event.

Thirty-seven quartz deformation lamellae were measured in 150 grains from the west side of the outcrop in the Monkton Quartzite (station S9, figure 8). The deduced positions are very similar to the stress positions deduced for the deformation lamellae at Shelburne Access Area, and hence the two are considered coeval (compare figures 6 and 9). The lamellae at S9 are thought to be older than the normal faults since horizontal slickensides indicating strike slip displacement are not present. If these faults were genetically related to the deformation lamellae then all but one should show right lateral displacement. According to figure 1 the Monkton-Winooski contact should be offset in a

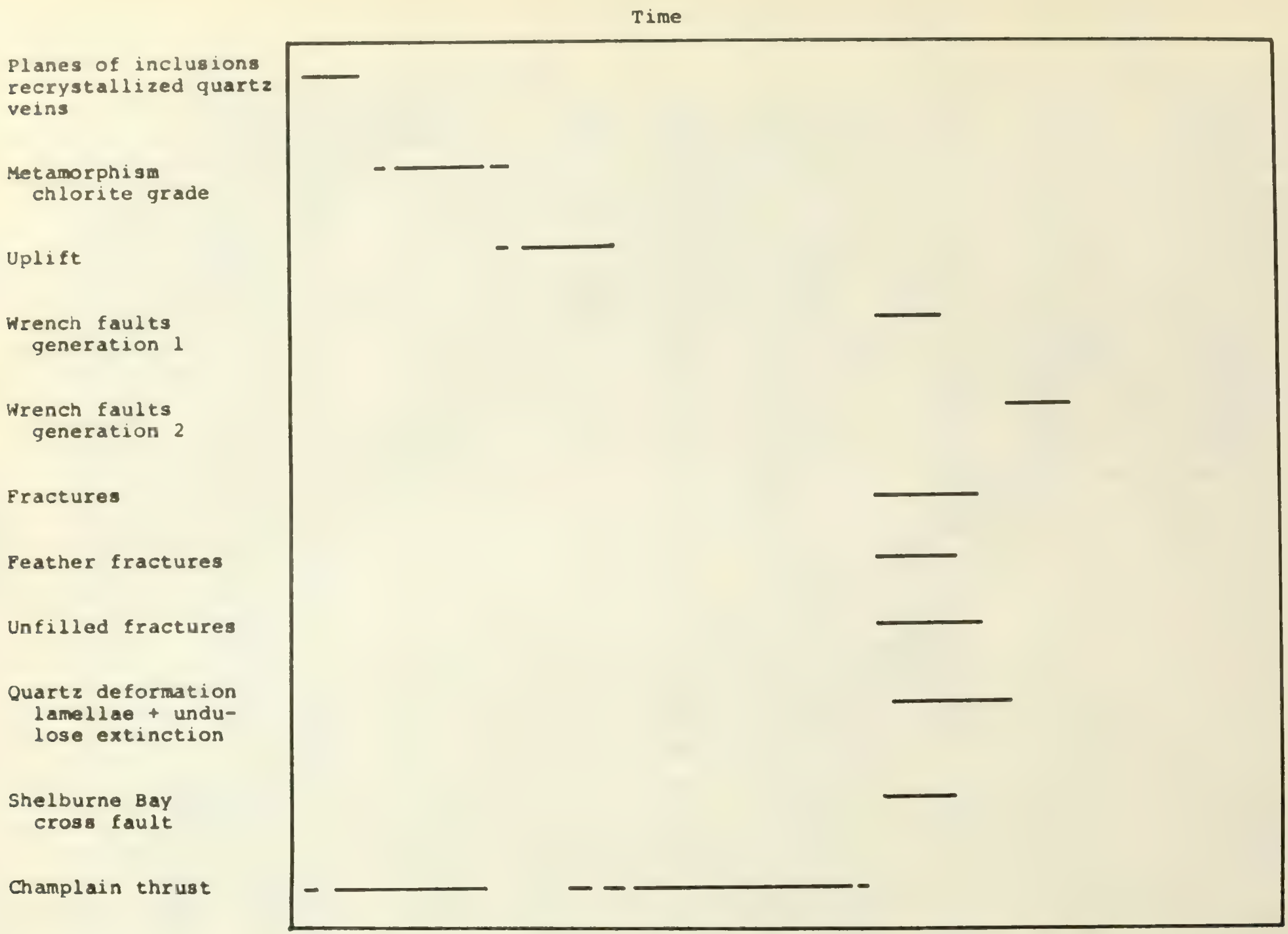

Figure 7. Chronology of structural events at Shelburne Access area Shelburne Bay, western Vermont.

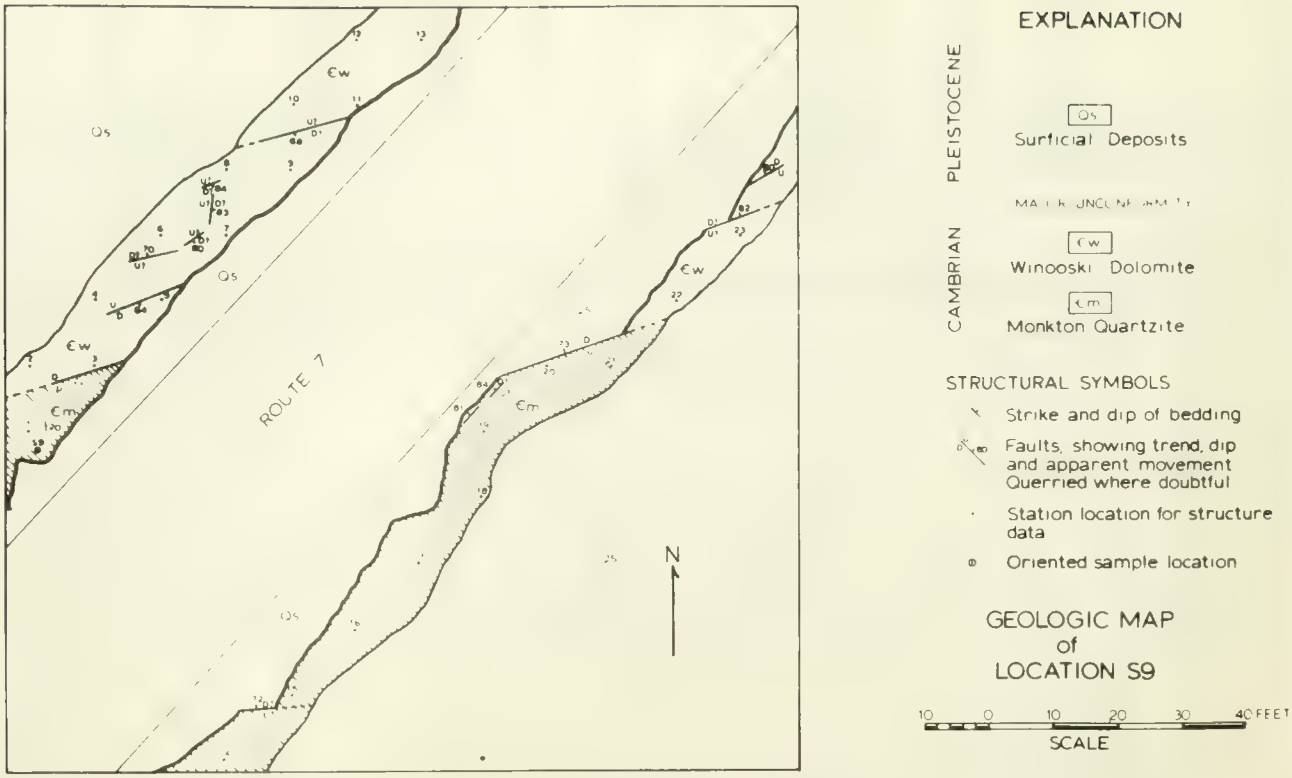

Figure 8. Geologic map of the normal faults at S9 just north of Shelburne, Vermont.

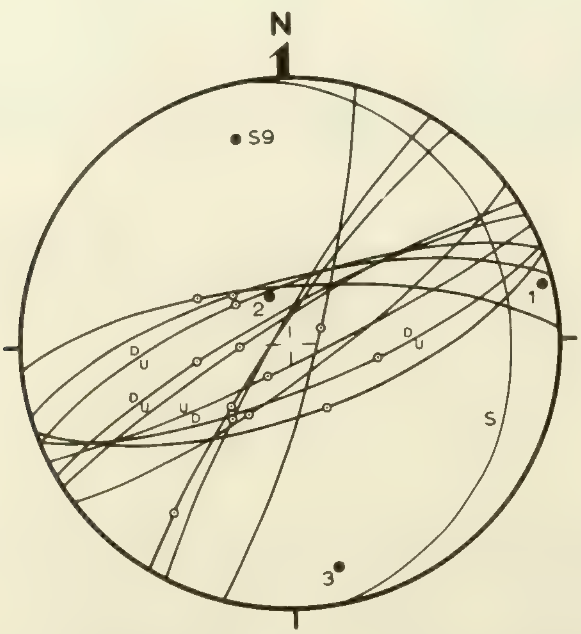

Figure 9. Lower hemisphere equal area projection showing the normal faults and associated slickensides at S9. The principal stress positions (1,2,3) deduced from quartz deformation lamellae in the Monkton Quartzite west of Route 7 are represented by solid dots. The generalized bedding at locality S9 (see figure 8) is shown by the great circle labelled S.

left lateral sense due to movement on the normal faults at this stop (figure 8).

In summary the structural chronology at S9 begins with the quartz lamellae which developed as a result of east-west compression associated with the generation 1 wrench faults at Shelburne Bay. Northwest-southeast extension then produced the normal faults which dominate the outcrop. This stress configuration is reflected in quartz lamellae from an outcrop 2 miles to the south of S9.

Stop 4. Pease Mountain near Charlotte (8, 10, figure 1) - The Champlain thrust and associated minor thrusts in its lower plate are well exposed on Pease Mountain (figure 10). The area was mapped by Cady (1945) and later remapped in greater detail by Welby (1961) and discovered outcrops of Bridport Dolomite on the western peak of the mountain. The area was mapped in still greater detail by students in field geology at the University of Vermont in 1969 and 1970. Our work has shown that the Bridport consists of two thrust slivers that have been subsequently deformed so that the bounding thrusts are systematically folded.

Stratigraphy: The abbreviated section at Pease Mountain includes part of the lower and upper members of the Monkton Quartzite which forms the upper plate of the Champlain thrust underlying the top and eastern slopes of the mountain. The Monkton is thrust on an overturned Middle Ordovician section that includes the upper part of the Glen Falls Formation, the Stony Point Formation, and the Iberville Shale. Slivers of Bridport Dolomite, a member of the Chipman Formation of Lower Ordovician age, are mapped along the western peak of the mountain. A few primary structures in the Bridport show that it is generally right side up.

Structure: Although thrusts dominate the structure on Pease Mountain, cleavage, folds and high angle faults are important aspects of the area.

Cleavage dips gently to the east in the shaly rocks of the Monkton Quartzite, the Bridport Dolomite, and the shales of middle Ordovician age. In the Monkton Quartzite the cleavage dips more steeply than the bedding which is a common relationship along the Champlain thrust (diagram A, figure 12).

Asymmetrical folds that deform the cleavage are restricted to the lower member of the Monkton Quartzite near the Champlain thrust. Six of these folds define a 65 degree separation arc with a deduced slip line that indicates movement of the upper plate in a N75W direction (diagram B, figure 12).

In the Iberville Formation on the east side of Route 7 (figure 10) two generations of folds are well developed and are

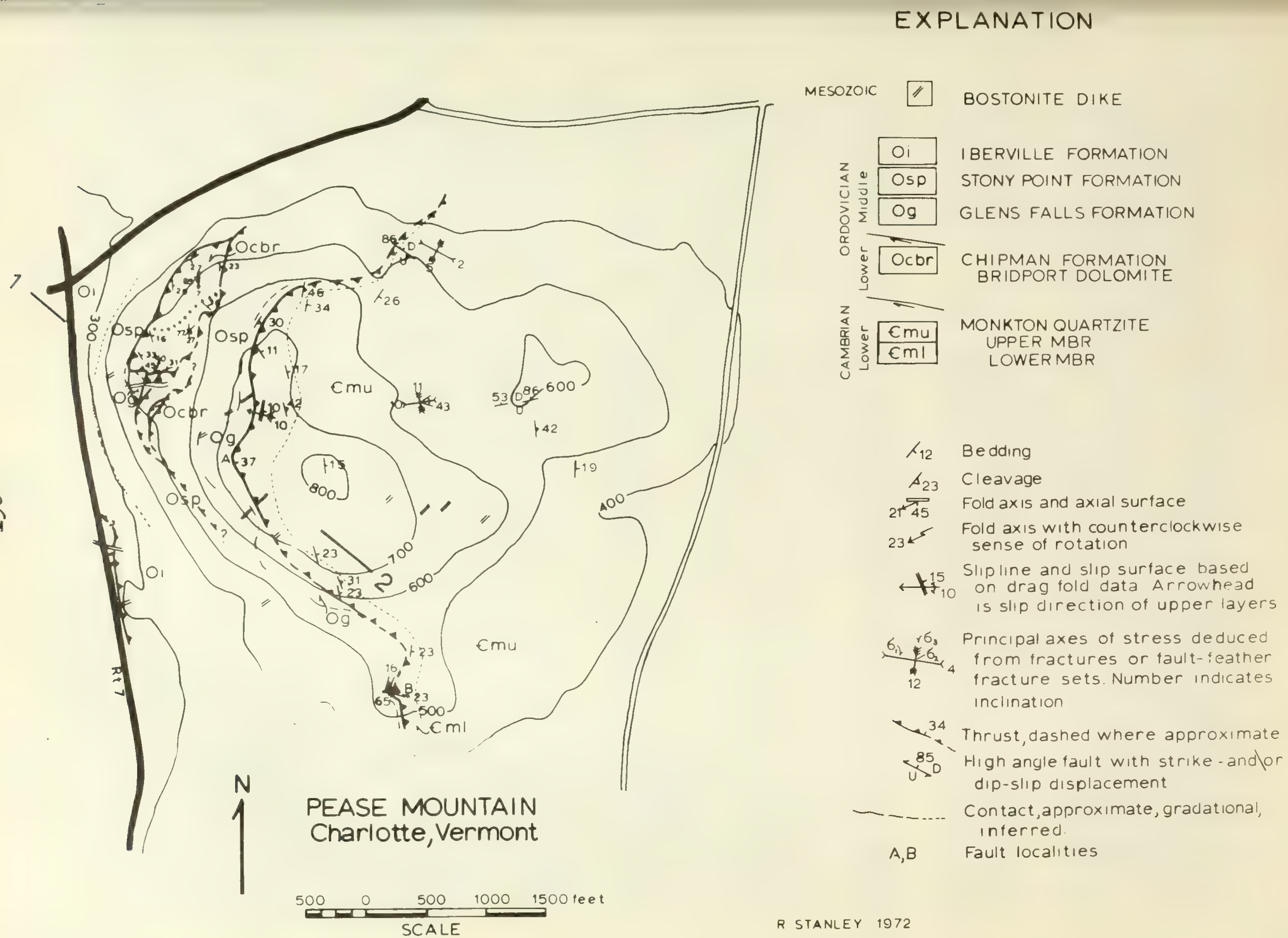

Figure 10. Geologic map of Pease 'lountain just east of Charlotte, western Vermont.

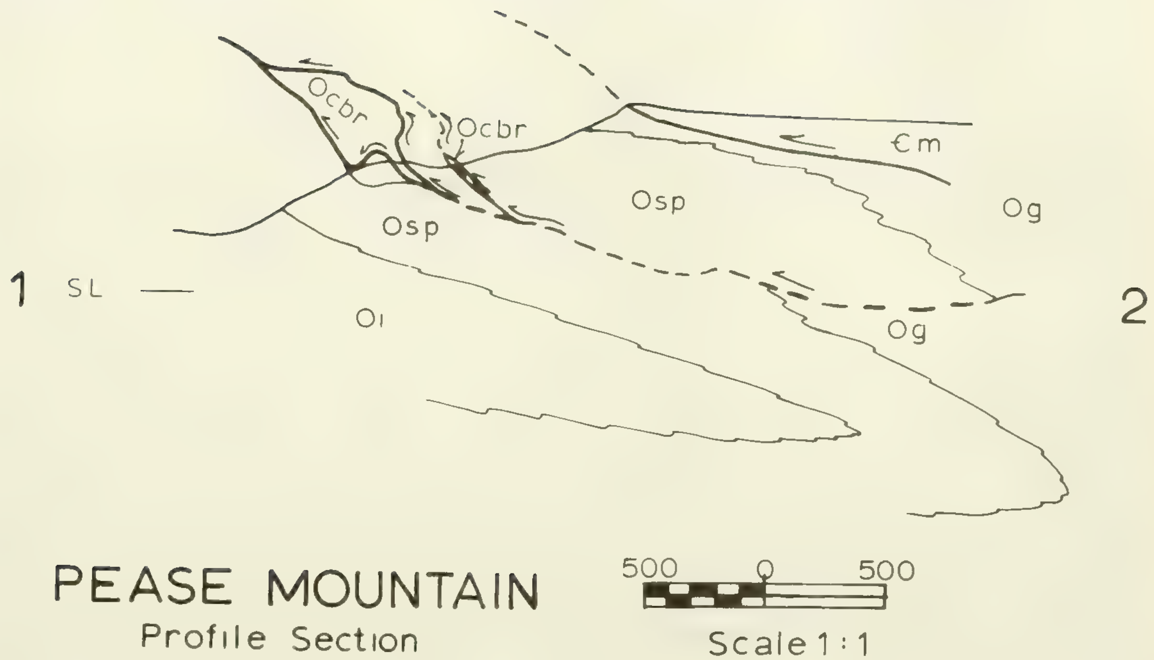

Figure 11. Modified profile section of Pease Mountain along a line of section labelled 1-2 on figure 10. No vertical or horizontal exaggeration. Displacement arrows only represent a component of the true direction of movement which is indicated by the slip direction deduced from the drag fold data shown in figure 10 and diagram B of figure 12.

Figure 12. Lower hemisphere equal area projections of selected structures on Pease Mountain. Diagram A shows poles to bedding and cleavage in the Monkton Quartzite. The intersection of bedding and cleavage and their anticlinal sense are represented by a solid dot with a concentric arrow. Diagram B shows the orientation of six drag folds which deform cleavage in the lower member of the Monkton Quartzite directly above the Champlain thrust north of point A on figure 11. Diagram C shows the orientation of generation one and two folds and the thrust fault in the Iberville Formation just east of Route 7 on figure 11. Diagrams D and E represent poles to bedding in the Bridport Dolomite. Dashed great circle represents the great circle that best approximates the distribution of poles. Diagram D contains 32 poles and diagram E contains 22 poles.

similar to the folds in the Stony Point Formation directly below the Champlain thrust at Lone Rock Point. At Pease Mountain the older folds are far more abundant than the younger folds which are only developed below a fairly continuous thrust at the south end of the outcrop. The orientation of each of these generations and the thrust is shown in diagram C of figure 12. The senses of rotation of the folds in both generations indicate movement to the northwest of upper beds over lower beds with a more northerly direction for the older set of folds.

The Champlain thrust is exposed at two localities (A and B, figure 10) where the lower member of the Monkton overlies highly deformed shaly limestones of the Glens Falls Formation. Silicified minor faults are common in the Monkton just above the thrust and suggest an earlier deformation perhaps associated with early movements on the Champlain thrust.

The Bridport slivers: The stratigraphic gap between the Bridport Dolomite and the surrounding Stony Point Formation leaves little doubt that the Bridport is bound by thrusts on the western part of Pease Mountain. Although the actual thrust surfaces are covered the systematic change in orientation of bedding in the dolostones and limestone of the Bridport near the thrust throughout the sliver and in the limestone and shale at the southern end below the thrust indicates that the bounding thrusts are systematically folded which is best seen around the southern end of the larger sliver. Poles to bedding in the Bridport define two diffuse great circles whose π pole (β point) plunges S36E at 25 degrees for the northern part and N40E at 25 degrees for the southern part (diagrams D and E, figure 12). Since there is no evidence supporting superposition of one of these folds on the other, it is concluded that the fold axis in the Bridport sliver curves through 80 degrees from the southern end of the sliver to the northern end.

The folded shape of the Bridport sliver indicates that it was systematically deformed after it was emplaced. It is suggested that the sliver was formed during the early stages of movement on the Champlain thrust and then was folded during subsequent movement on the thrust.

Stop 5. Mount Philo near Ferrisburg (7, figure 1) - Mount Philo is located along the Champlain thrust on the north limb of the Monkton culmination (Cady, 1945; Doll, and others, 1961) just south of Charlotte, Vermont (figure 1). Although the Champlain thrust is not exposed in this area, numerous east-west faults, folds, and several thrusts are well developed in the Monkton Quartzite which forms the upper plate of the thrust. Five oriented specimens of quartz deformation lamellae were analyzed by Sarkisian (1970) from three separate localities (S11, S2, SF) located in figure 13.

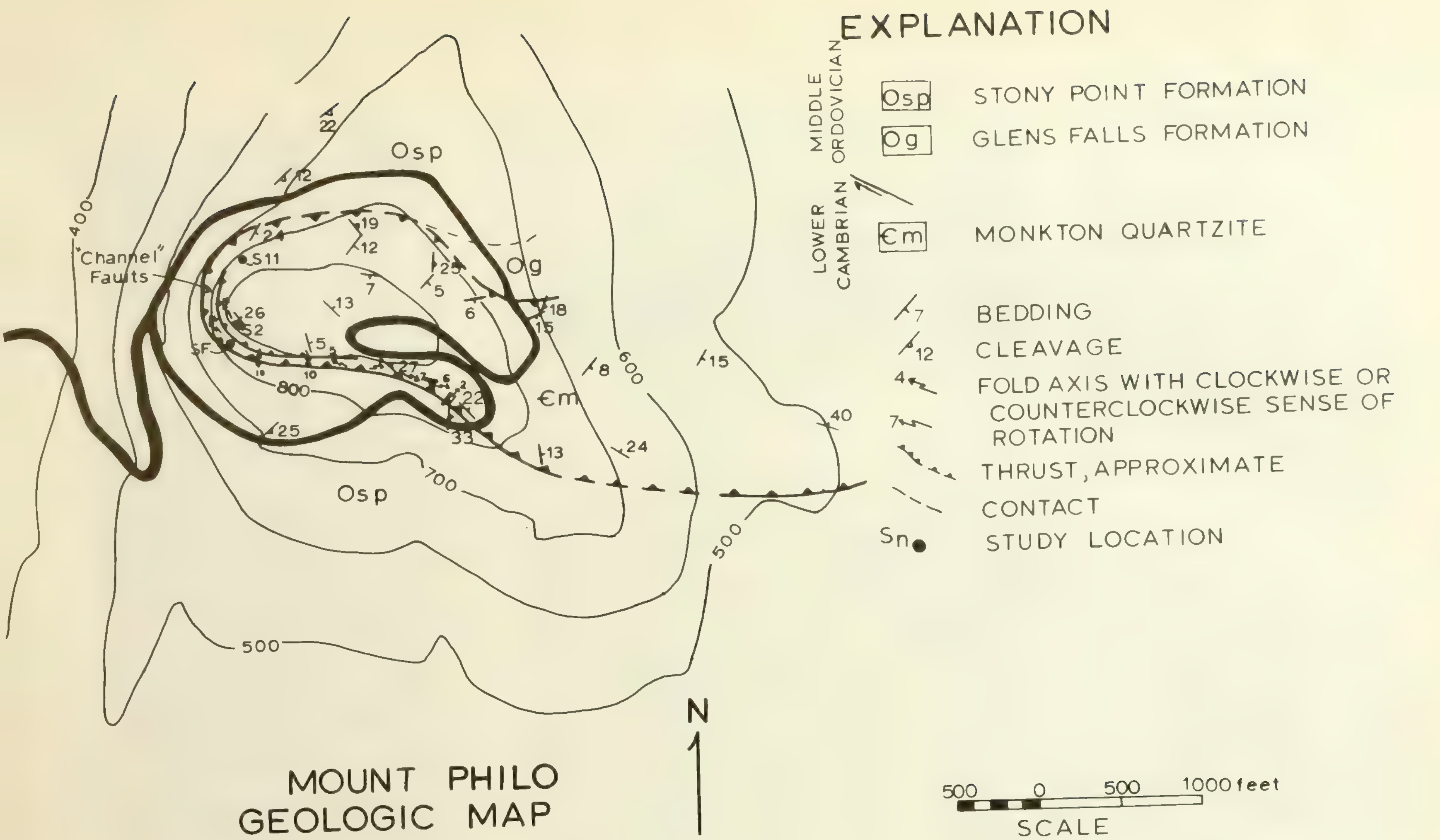

Figure 13. Geologic map of Mount Philo, North Ferrisburg, Vermont, near the southern boundary of the map in figure 1.

Megascopic Structures: Bedding, slaty cleavage, asymmetrical concentric folds, fractures, and faults are well displayed along the southern and western cliff of Mount Philo (figure 13). Crossbedding and ripple marks indicate that the Monkton Quartzite is right-side-up throughout the area. Cleavage is well developed in the thin shaly beds of the Monkton and dips eastward more steeply than the bedding where it is not folded. Large asymmetrical folds in the southern and western cliffs deform the cleavage and plunge at very gentle angles in various directions. The sense of rotation of 12 of these folds on the upper plate of the Mount Philo thrust locate a horizontal slip line that trends N55W and indicates movement of the upper beds northwestward (figure 14). Four fairly large folds are also present directly below the Mount Philo thrust and indicate a slip direction slightly south of east. On the western cliff of Mount Philo below the thrust high angle faults commonly dip northward and southward. Although movement on the surfaces are commonly normal, movement in the reverse sense was noted. In two key areas north-dipping faults high on the cliff flatten at a lower elevation and pass into high angle south-dipping faults further on and up the cliff. These fault surfaces, therefore, form U-shaped channels and show either normal or reverse movements across the fault surface. The Mount Philo thrust cut these high angle faults and hence is younger in age.

The Mount Philo thrust crops out for at least 700 feet along the southern and western cliffs of Mount Philo (figure 13). It is a sharp, undeformed surface that dips gently eastward and truncates the asymmetrical folds within the lower portion of the Monkton Quartzite.

Several fracture sets are well developed on Mount Philo. They cut the folds, high angle faults, and the Mount Philo thrust. At locality S2 (figure 15) Sarkisian measured 163 fractures across one of the asymmetrical folds. The resulting fabric (figure 15) shows three statistical fracture sets which correspond to the maxima in the contoured equal area diagram. This fabric is undeformed by the fold since separate plots on opposite limbs of the fold are similar to the diagram in figure 15. The plane of symmetry bisecting fractures 1 and 3 is perpendicular to fracture 2 and is approximately parallel to the slip line determined from the asymmetrical folds.

Microscopic Structures: Five oriented samples of Monkton Quartzite were collected from three separate localities on the western side of Mount Philo (S11, S2, SF, figure 13). One sample (S11) was collected below one of the east-west faults, another (S2) comes from the locality where 163 fractures of figure 15 were analyzed and the remaining three (S10, S12, S13 at locality SF) were collected from the limbs and hinges of an asymmetrical

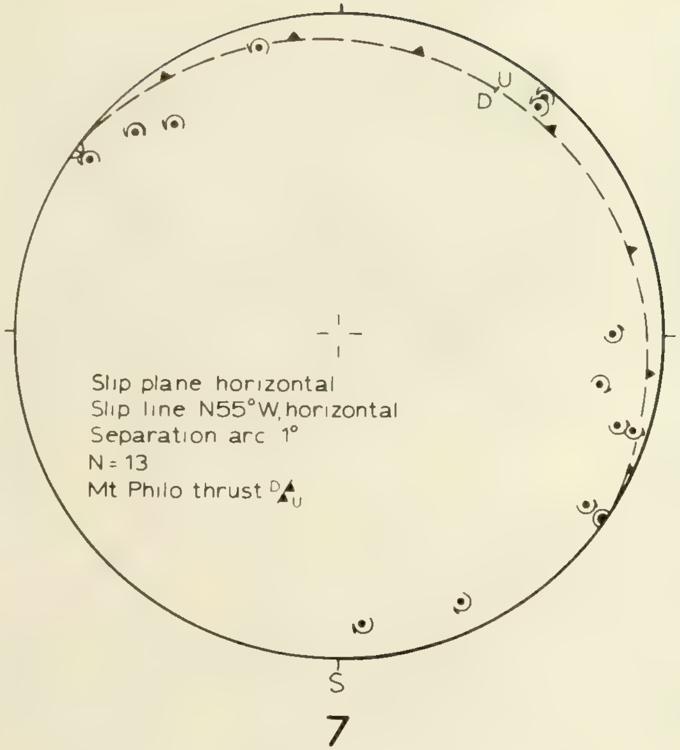

Figure 14. Lower hemisphere equal area projection of 13 drag folds that deform cleavage in the lower member of the Monkton Quartzite just above the Champlain thrust on Mount Philo. The orientation of fold axes and their respective sense of rotation are shown by solid dots and concentric arrows. The number 7 refers to diagram 7 in figure 2.

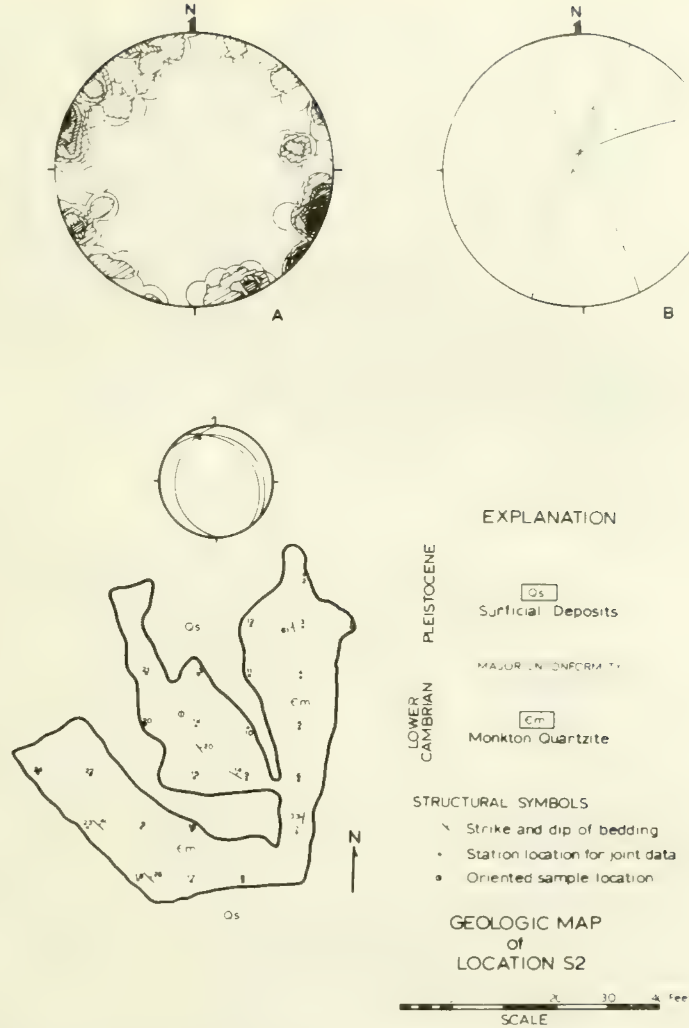

Figure 15. Geologic map and lower hemisphere equal area projections of fractures in a syncline in the upper member of the Monkton Quartzite at locality S2 in figure 13. The orientation of the synclinal fold axes is shown in the small equal area projection.

drag fold. Oriented samples were, therefore, selected from all of the megascopic structures except the Mount Philo thrust.

For each sample 75 (50 for S11) quartz grains were studied from each of three mutually perpendicular thin sections. The quartz lamellae are similar to those described for the Shelburne Access Area and locality S9.

Synoptic diagram A in figure 16 shows the principal stress positions deduced from deformation lamellae on Mount Philo. For all the samples the poles to lamellae define small circle girdles with radii that range from 55 to 64 degrees. This pattern corresponds to a cone of lamellae oriented less than 45 degrees to the central cone axis, σ_1 . The principal stress directions deduced from the five specimens are remarkably constant in orientation and configuration. The σ_1 directions fall in a narrow 30 degree arc oriented north of west (average direction, N75W). The σ_2 and σ_3 directions are approximately equal in value and hence, define the plane perpendicular to σ_1 . The trend of σ_1 is approximately 20 degrees counterclockwise to the trend of the slip direction (N55W) deduced from the asymmetrical drag folds (figure 16).

Since the quartz fabric axes have not been rotated by the folds at SF, the quartz deformation lamellae were superposed on this fold after it had fully developed. East-west fractures in sample S11 collected near the high-angle faults offset the deformation lamellae and suggest that the lamellae are older than the Mount Philo thrust and its associated channel faults.

Relationship of Quartz Deformation Lamellae to the Megascopic Structures: The quartz deformation lamellae on Mount Philo have resulted from a nearly horizontal maximum compressive stress generally oriented in N75W direction. The values of σ_2 and σ_3 were approximately equal during lamellae development. The quartz lamellae reflect a stress configuration that is compatible with the north-trending Champlain thrust. It is also similar to the stress configurations deduced from samples M3 and M4 at Shelburne Access Area which are in turn correlated with the first generation wrench faults and the Shelburne Bay cross fault. Thus the lamellae at Mount Philo probably developed with the younger wrench faults which cut the Champlain thrust.

At Mount Philo the deformation lamellae are younger than the asymmetrical drag folds since the lamellae fabric axes remain constant in orientation across the fold. In sample S11 small shear fractures offset quartz deformation lamellae. These fractures parallel the east-west channel fault and hence are considered younger than the deformation lamellae. This temporal relationship would further support the conclusion that the high

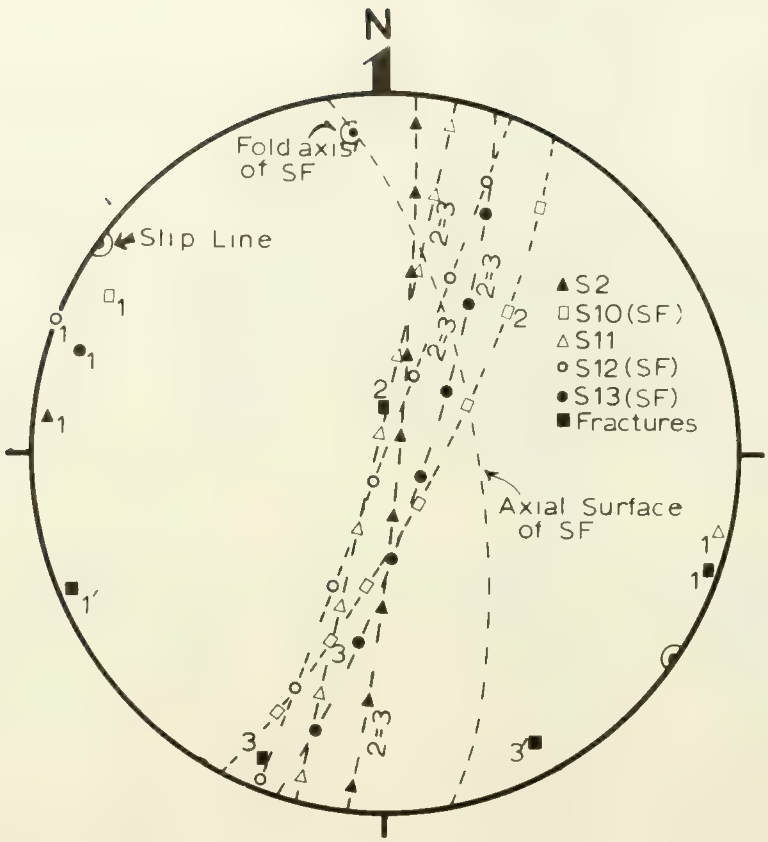

Figure 16. Synoptic diagram of principal stress positions deduced from quartz deformation lamellae (S2, S10, S11, S12, S13) and fractures. The numbers 1, 2, 3 correspond to the principal compressive stresses with 1 representing the direction of maximum compressive stress. Alternative principal stress positions for fractures are represented by primed and unprimed numbers. The slip line deduced from the drag folds in figure 14 is also included in the projection.

angle faults and the Mount Philo thrust are younger than the asymmetrical folds.

In summary the structural sequence at Mount Philo begins with the development of cleavage and is followed by the folding of the Monkton into west-facing folds possibly as a result of movement of the Champlain thrust to the northwest. Subsequent deformation produced the quartz deformation lamellae which are thought to be coeval with the first generation of wrench faults at Shelburne Bay. Continued west northwest - east southeast compression resulted in the channel faults and the Mount Philo thrust. Fracturing subsequently developed and may reflect a change in orientation of the principal stresses although the fracture can be related to the previous stress configuration.

Summary of structural chronology. The temporal relationship among the structures at the five localities along the Champlain thrust are summarized in figure 17. Reasons supporting their age assignments are discussed at each locality and will not be repeated here. The following comments will be restricted to the relationship of these structures to such major structures or events as the Champlain thrust, Hinesburg synclinorium, Hinesburg thrust, and the various orogenies known in the Appalachians.

As shown in figure 17. and emphasized at different localities, the Champlain thrust is thought to have undergone a multiple history beginning with initial emplacement during the Taconic orogeny and ending with renewed movement in a subsequent orogeny, probably the Acadian of Middle Devonian age. Since the youngest rocks below the Champlain thrust are Middle Ordovician in age it seems unjustified to restrict its development to the Acadian orogeny as suggested by Cady (1969, p. 75). Subsequent movement apparently did occur during the Acadian or possibly the Allegheny orogeny since the chlorite-grade rocks of the upper plate now rest on essentially unmetamorphosed rocks of the lower plate. Radiometric work in the northern Taconics (Harper, 1968), along the Sutton-Green Mountain anticlinorium (Cady, 1969, p.104), and in Quebec (Rickard, 1965) indicate that recrystallization in northwestern Vermont was older than 400 m.y. and hence of Taconic age. Petrologic work by Albee (1968) lends further support to this conclusion. Thus renewed movement on the Champlain thrust is restricted to post Taconic activity.

Based on the foregoing conclusions the second generation of younger folds in the lower plate, the asymmetrical folds which deform cleavage in the upper plate, and the deformation of the Bridport sliver on Pease Mountain are contemporaneous with renewed activity on the Champlain thrust. The older generation of folds in the lower plate, the original emplacement of the Bridport sliver, the formation of cleavage in the Monkton, and low grade metamorphism may well be associated with, or just after,

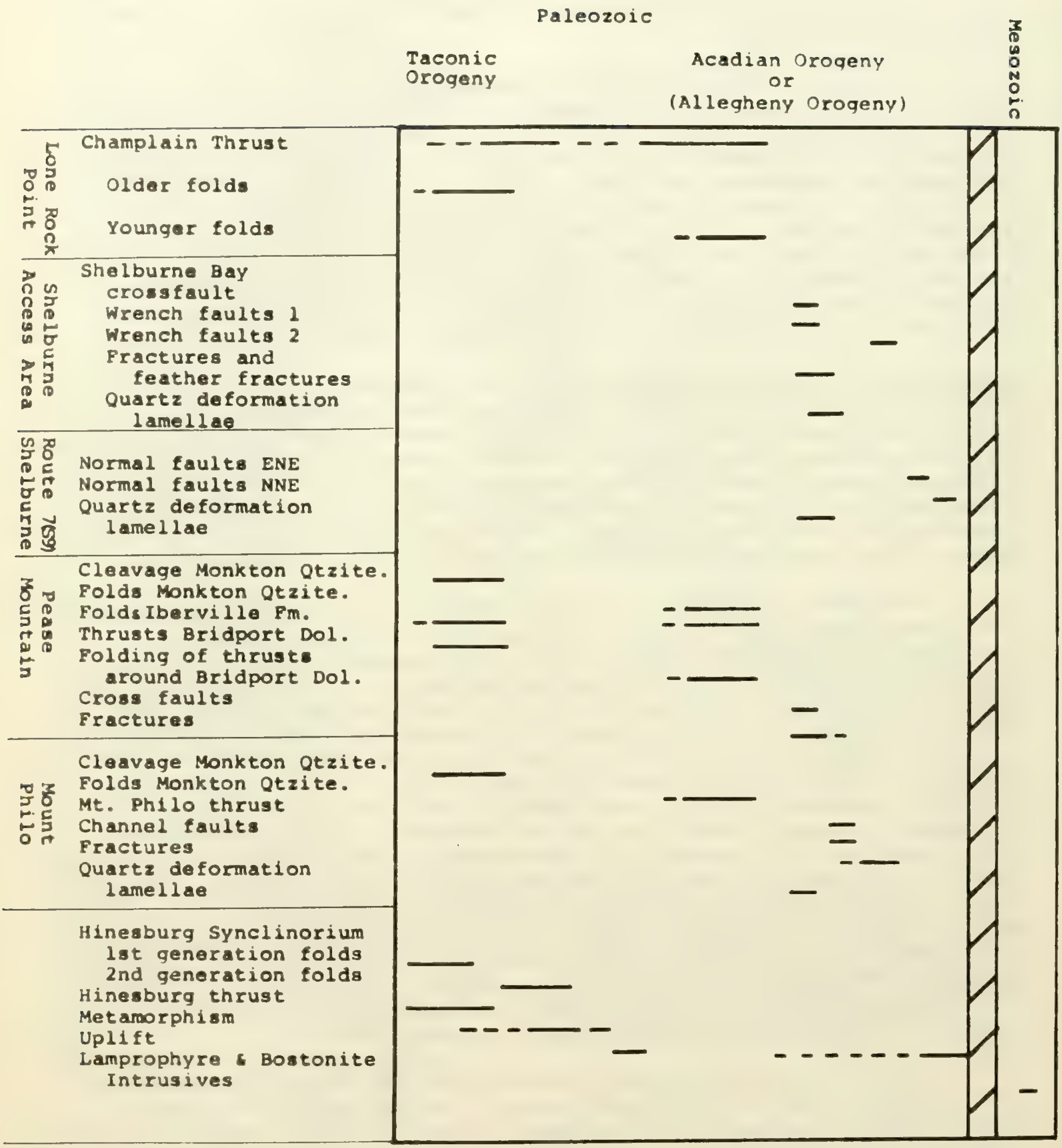

Figure 17. Chronology of selected structural events in the Hinesburg synclinorium and along the Champlain thrust in the central part of western Vermont.

early development of the thrust. The two generations of wrench faults, the normal faults, and the quartz deformation lamellae are younger than the Champlain thrust and are probably Acadian in age although an Allegheny age is certainly possible.

The structures in the Hinesburg synclinorium and along the Hinesburg thrust can be placed in this chronological sequence although our work is still in progress (Gillespie and others, this guidebook). The rocks in the southern part of the synclinorium have been involved in at least two, and in some places, three generations of folds. The axes of the first generation of tight folds plunges gently southeastward with a well-developed closely spaced cleavage. These folds are, in turn, folded into rather open folds with north plunging axes and steep eastward dipping axial surfaces. The map pattern in the southern part of the Hinesburg synclinorium is actually a product of both of these fold events (figure 1, Doll and others, 1961). Analysis of quartz lamellae in the Monkton at Mount Philo, S8 and farther east by Sarkisian and Marcotte indicate the lamellae are younger than the second generation of folds since the deduced stress positions are not deflected by the major folds. Thus the formation of the Hinesburg synclinorium predates the wrench faults and associated quartz deformation lamellae. Since there has been recrystallization of micaceous material in the axial surfaces of the first and possibly the second generation of folds, these events are probably Taconic.

The Hinesburg thrust is clearly folded by the second generation of folds in the synclinorium and therefore is also considered to be Taconic in age.

The lamprophyre and bostonite dikes that cut the Champlain thrust, the Hinesburg synclinorium and the upper plate of the Hinesburg thrust are the youngest structures recognized in west-central Vermont. These intrusives are Mesozoic in age since K-Ar measurements on biotite from the syenite stock at Barber Hill in Charlotte indicate an age of 111 + or - 2 m.y. (Armstrong and Stump, 1971). Similar work on a lamprophyre from Grand Isle yield an age of 136 + or - 7 m.y. (Zartman and others, 1967).

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Trip B-6

SEDIMENTARY CHARACTERISTICS AND TECTONIC DEFORMATION OF MIDDLE AND UPPER ORDOVICIAN SHALES OF NORTHWESTERN VERMONT NORTH OF MALLETTS BAY

by

David Hawley, Hamilton College, Clinton, New York

Introduction

The central lowland of the Champlain Valley is underlain by Cambrian and Ordovician sedimentary rocks, bordered on the west by the Adirondack Mountains of Precambrian crystalline rock upon which Cambrian sandstone lies unconformably, and against which sedimentary rocks have been dropped along normal faults. The lowland is bordered on the east by low-angle thrust faults on which massive dolomite, quartzite, and limestone, as old as Lower Cambrian, from the east over-rode weaker Ordovician shale and limestone. The westernmost thrusts, the Highgate Springs thrust in the north, and the overlapping Champlain thrust in the south, trace an irregular line a few feet to 3 1/2 miles inland from the east shore of Lake Champlain. For most of the distance between Burlington and the Canadian border, the high line of bluffs marking the trace of the Champlain Thrust are composed of the massive, Lower Cambrian Dunham dolomite.

The shales, youngest rocks of the autochthonous lowland sequence, outcrop on most of the islands in Vermont, and the mainland between the thrusts and the lake. Although exposures on almost continuous shore-line bluffs are excellent, there are few outcrops inland because of glacial cover and low resistance of the shales to weathering. Fossils are rare in the older calcareous shale (Stony Point) and absent in the younger non-calcareous shale (Iberville). The lithic sequence was established almost entirely on structural criteria. Where it can be found, the Hathaway submarine slide breccia structurally overlies the Iberville.

Description of Formations

Glens Falls Limestone

Kay (1937, p. 262-263) named the lower Glens Falls the Larrabee member, found it to be 72 feet thick on the Lake Champlain shore in the north-western part of South Hero Township, Vermont, and to be composed there of

thin-bedded, somewhat shaly limestone. Fisher (1968, p. 27) has found the Larrabee member to be 20 to 30 feet thick in the vicinity of Chazy, N.Y., and to be coarse-grained, medium- to thick-bedded light gray limestone full of fossil debris (brachiopods, crinoids, pelecypods, and trilobites).

The upper Glens Falls was named the Shoreham member by Kay (1937, p. 264-265), and described as the zone of Cryptolithus tesselatus Green, a distinctive trilobite. He found 30 feet of the Shoreham exposed in the lakeshore in northwestern South Hero Township. Fisher (1968, p. 28) prefers to call this the Montreal limestone member, following Clark's usage for the Montreal area (1952), and has described it as medium dark gray to dark gray argillaceous limestone with shale partings, fossiliferous with trilobites, brachiopods, molluscs, and bryozoa. He estimates it to be 150-200 feet thick in Clinton County, N.Y.

Cumberland Head Formation

The "Cumberland head shales" was a term used, but not carefully defined by Cushing (1905, p. 375), referring to the interbedded shale and limestone forming a gradation between the Glens Falls and the overlying Trentonian black shales. Kay (1937, p. 274) defined it as "the argillaceous limestones and limestone-bearing black shales succeeding the lowest Sherman Fall Shoreham limestone and underlying the Stony Point black shale." He measured 145 feet on the west shore of South Hero Island, Vt., just south of the Grand Isle-South Hero town line. The lower 30 feet have 8- to 12-inch beds of gray argillaceous limestone interbedded with dark gray calcareous shale. Above that the shale is predominant, but limestone beds are abundant, 3 to 12 inches thick with undulating surfaces, interbedded with half-inch to 12-inch layers of black calcareous shale. Less than one third of the Cumberland Head has more than 50 per cent shale, and about half has more than 60 per cent limestone beds. Some units as thick as 15 feet have 80 per cent limestone beds. The proportion of shale increases gradually but not uniformly upward.

Stony Point Formation

The Stony Point shale was defined by Ruedemann (1921, p. 112-115) as "hard, splintery dark bluish-gray calcareous shale" at Stony Point, 1 1/2 miles south of Rouses Point, N.Y., on the west shore of Lake Champlain, and correlated faunally with upper Canajoharie shale of the Mohawk Valley (Middle Trentonian).

The base of the Stony Point is exposed on the lake shore 0.55 miles south of the breakwater at Gordon Landing, the eastern end of the Grand Isle-Cumberland Head ferry. Deposition was continuous from the Cumberland Head up into the Stony Point, and the contact is somewhat arbitrarily chosen where the proportion of shale increases upward, and the wavy, irregular limestone bedding of the Cumberland Head gives way upward to smooth, even limestone beds of the Stony Point. The 215 feet of Stony

Point formation exposed here is interbedded dark gray calcareous shale with light-olive-gray weathering, dark gray fine-grained limestone in beds of 1 to 12 inches, about 70 per cent shale. Two units about 9 feet thick are about 80 per cent limestone beds.

The thickest and least deformed measurable section of Stony Point begins 0.6 mile north of Wilcox Bay and extends for 1.8 miles northward along the shoreline bluffs of northwestern Grand Isle (Hawley, 1957, p. 59, 87-89). In this section of 635 feet, there are a few gross vertical lithic variations which are recognizable throughout this field area. Above the lower 215 feet, as described above, the percentage of calcareous shale decreases upward. Olive-gray to light-olive-gray weathering, dark gray argillaceous limestone appears in increasing proportion through the upper 400 feet of this section, where the percentages are: argillaceous limestone, commonly silty, 66 per cent; calcareous shale, 29 per cent; fine-grained limestone beds, 5 per cent.

The argillaceous limestone commonly occurs in thin, even beds (one quarter to three quarters of an inch) with fine lighter-and darker-gray laminae, but occasional beds reach 10 inches. Thicker-bedded zones suggest cyclic deposition: from calcareous shale (1 to 4 inches) upward through 5 to 6 inches of laminated argillaceous limestone, to a 1- to 3-inch bed of fine-grained limestone; then through 4 to 5 inches of argillaceous limestone to 1 to 4 inches of calcareous shale. Where the interval between calcareous shale beds is thinner, the fine-grained limestone bed in the middle is missing. The proportion of silt and argillaceous material in harder argillaceous limestone varies greatly. Intricate, fine, current cross-bedding occurs in four thin zones, indicating currents flowing northeastward.

Above this zone rich in laminated argillaceous limestone the proportion of calcareous shale increases, and 239 feet near the top of the Stony Point is composed entirely of calcareous shale. This shale section, 1.4 miles S 37° W from Long Point, North Hero, Vt., is assumed to represent the uppermost part of the Stony Point because it lies on the nose of a long, northeastward-plunging anticline between a thick argillaceous limestone section to the southwest, and a large area of Iberville shale to the north and northeast.

In this field area it is not possible to measure the entire thickness of the Stony Point, but from piecing together several measurable sections a minimum thickness is 874 feet. The total thickness is estimated to be 1000-1500 feet, allowing for probable thicknesses that could not be measured in the middle and upper parts of the Stony Point (Hawley, 1957, p. 83). In the log of the Senigon well near Noyan, Quebec, about 4 miles north of the international boundary at Alburg, shale apparently equivalent to the Stony Point is 924 feet thick (Clark and Strachan, 1955, p. 687-689).

Iberville Formation

The Iberville formation was named by Clark (1934, p. 5) for its wide outcrop belt in Iberville County, southern Quebec, about 10 miles north of

the international boundary at Alburg, Vt. Clark (1939, p. 582) estimated the Iberville to be 1000-2000 feet thick in its type area.

The base of the Iberville has a gradational contact and was chosen on the basis of lithic criteria by which it can be most easily distinguished from the Stony Point. The Stony Point is entirely calcareous shale and argillaceous limestone with occasional beds of light-olive-gray weathering, dark gray fine-grained limestone. Above the lower transition section, the Iberville is composed of interbeds of medium to dark gray noncalcareous shale (1-12 inches, usually 2-4 inches), moderate-yellowish-brown weathering, dark gray laminated dolomitic siltstone (one quarter inch to 10 inches, usually 1/2- 1 1/2 inches), and occasionally moderate-yellowish-brown weathering, dark gray fine-grained dolomite. The most conspicuous change from the Stony Point is the appearance of the yellowish-brown weathering dolomite beds, and the noncalcareous shale which is more brittle and more lustrous on cleavage surfaces than the calcareous shale. The transition section is at least 72 feet thick at Appletree Point in northern Burlington (Hawley, 1957, p. 64), and may be as thick as 200 feet. A section from Stony Point to Iberville is almost continuously exposed, though somewhat deformed, along the lakeshore southeastward for a half mile from Kibbee Point, in northeastern South Hero Township, Vt. The base of the Iberville is defined as the first appearance of the noncalcareous shale and dolomite beds.

Iberville shale and dolomitic siltstone show remarkable rhythmic bedding. The base of each cycle is a sharp contact, sometimes a slightly scoured surface, on which a thin bed (0.25-0.75 inch) of yellowish-brown weathering, dark gray laminated dolomitic siltstone was deposited. The typical siltstone layer becomes finer-grained upward with decreasing quartz and dolomite, and increasing argillaceous and carbonaceous material, and grades into dark-gray noncalcareous thin-cleaving shale (1-4 inches). Usually at the top is an eighth to three quarters of an inch of grayish-black shale interlaminated with the dark gray. Occasionally the dolomitic siltstone may be missing at the bottom of the cycle, or the grayish-black shale laminae missing at the top. Ripple-drift cross-lamination is a common feature of the dolomitic siltstone layers. In some beds only a single story of ripples were built, but in others down-current ripple drift continued long enough to form two, and occasionally three or four tiered beds. Current directions indicated by the ripple cross-lamination are invariably southwestward in the Iberville, in contrast to northeastward in the Stony Point.

Six thicker (5-10 inches) non-laminated graded siltstone beds with 1 mm.-long shale flakes in their lower parts are found on northeastern Burton Island, southwest of St. Albans Point. They grade finer upward, and some are laminated above the lower third. One has large (5 by 1 1/4 inches is the largest) angular shale fragments in the mid-portion. They commonly have contorted lamination in the middle, above which lamination is more marked, and they are topped with drift ripples grading upward into shale.

The thickest measurable sections of the Iberville are 732 feet, with an estimated 2200 depositional cycles, on the west side of Woods Island,

and 304 feet with an estimated 1215 cycles on Clark Point, southwestern Hog Island, West Swanton, Vt. The cyclic character of the Iberville layers, the graded beds, graded laminated beds, and convolute laminae, are all characteristic of sedimentation by turbidity currents (Kuenen, 1953; Bouma, 1962, p. 48-54).

Hathaway Formation

The Hathaway formation, named for Hathaway Point on southeastern St. Albans Point, Vt. (Hawley, 1957, p. 68), designates argillite and bedded radiolarian chert, commonly intensely deformed, with included small fragments to large blocks of quartz sandstone, coarse graywacke, dolomite, limestone, and chert. Some fragments strongly resemble dolomite and dolomitic siltstone beds of the underlying Iberville, but the coarse sandstone, chert and graywacke are unlike any strata in the autochthonous formations of the Champlain lowland. The graywacke resembles the earliest Cambrian Pinnacle Formation, which outcrops in a north-south trending area 8 to 10 miles east of northern Lake Champlain (Stone & Dennis, 1964, p. 19). Where the Hathaway and Iberville are in contact or close proximity, there is marked disparity in intensity and nature of their deformation. The Hathaway appears to have deformed by flowage without the development of good cleavage, commonly with disintegration of less mobile beds into blocks and boulders. The Iberville has undergone much less intense folding and faulting, of a type normally associated with the regional structure. For these reasons, the Hathaway is inferred to be a submarine slide breccia initially deformed while its muddy constituents were still soft.

The best accessible exposures of the Hathaway are on Hathaway Point, and extending north for 1200 feet from Beans Point on the east shore of the lake, in northwestern Milton Township, Vt. As fate would have it, the most impressive and extensive exposures of the Hathaway are on Butler Island, between St. Albans and North Hero, accessible only by boat. Almost all of Butler Island is composed of the Hathaway, which is usually a mashed, streaky light and dark gray argillite with inclusions of dolomite, dolomitic siltstone, and occasionally black chert and graywacke, from 1 by 2 to 8 by 24 inches. On the southeast side of Butler Island are found the largest inclusions in the Hathaway: blocks of dolomitic siltstone up to 3 by 20 feet, and coarse-grained graywacke up to 15 by 50 feet. Argillite foliation wraps around these blocks, and around innumerable smaller pebbles and boulders. Hawley has described in detail these and other localities (1957, p. 68-75).

Summary of Depositional History

The fossiliferous limestones of the Glens Falls and older formations in this area indicate rather shallow, clear-water carbonate deposition, often in an environment of considerable wave and current turbulence (reefs, coarse calcarenites, and cross-bedding in the upper Chazyan). In the Cumberland Head formation fossils are much scarcer and there is a

transition from the shallow water carbonate environment to a muddier, deeper water depositional environment. The lower two hundred feet of the Stony Point is 70 per cent calcareous shale, and the next 400 feet is laminated argillaceous limestone (66%) interbedded with calcareous shale (29%) and hard, purer fine-grained limestone (5%) in a somewhat cyclic pattern. Current cross-lamination indicates flow toward the northeast. The complete absence of primary structures associated with shallow water, and the fine lamination of the argillaceous limestone, and the paucity of fossils, suggest a deeper, quieter, muddier depositional environment.

Through the lower hundred feet (or more) of the Iberville, a marked change in the character of the rock appears with dolomite replacing limestone as the hard, fine-grained interbeds, and noncalcareous shale replacing the calcareous shale of the Stony Point. At some unknown distance above the base, a section of at least 730 feet shows cyclic interbedding of noncalcareous shale and graded, laminated dolomitic siltstone commonly with current cross-lamination. The currents flowed toward the southwest. This suggests the changed character of the rock is at least partly the result of a change from a westward source of sediment (for the Stony Point), to an eastward source for the Iberville, and that turbidity currents dominated the depositional character of the Iberville. Uplift of deep sea bottom east of the Champlain Valley in late Mohawkian and early Cincinnatian time could have provided the new source of sediment and the westward slope down which the turbidity currents flowed. Some simultaneous deepening of the Champlain Valley region also occurred.

The Hathaway formation, composed of argillite and bedded radiolarian chert, chaotically deformed, with included masses of limestone, dolomite, dolomitic quartz siltstone and sandstone, coarse graywacke, and chert, is interpreted as a submarine slide breccia. Some of the types of inclusions, particularly the graywacke and chert, are unknown in autochthonous underlying formations of the Champlain Valley and in regions to the south and west. The slide (or slides?) seem to have come from the east, down the slope suggested by the direction of flow of turbidity currents which deposited sediment in the Iberville. The Taconic orogeny was occurring at this time, and some believe that the major thrusts of western and northwestern Vermont accompanied this orogeny. If this be true, thrust fault escarpments on the sea bottom to the east of the Champlain Valley could account for the slides and the assemblage of inclusions in the Hathaway. Earthquakes associated with the Taconic orogeny may have triggered the turbidity currents of the Iberville.

Tectonic Deformation

The shales are complexly folded and sheared, with fold axes trending a little east of north in the southern part of the area, and swinging more toward the northeast (N 20° - 30° E) in the north. Although elongate narrow belts of intense deformation parallel fold trends, separated by broader belts of more gentle folding, general intensity of deformation increases toward the

Champlain and Highgate Springs thrusts. In areas underlain by shale, particularly in North Hero and Alburg, the topographic "grain" of long, low hills accurately reflects the trends of fold axes. From Grand Isle northward the smaller folds plunge northward and southward, but the pattern of structural elements and formational boundaries indicates the northeastward plunge is more prevalent and perhaps a bit steeper. The area might be visualized as having northeastward trending folds imposed on an eastward regional dip, though there are many individual exceptions to this generalized picture.

Fracture cleavage is nearly everywhere present in the more argillaceous beds of the Stony Point and Iberville formations. The term is used here as defined by Swanson (1941, p. 1247), "the structure is due to closely spaced planes of parting a certain small distance apart," and "as a rule it is possible to see that the rock between the planes of parting. . . has no structure parallel to them, or at most any parallel structure is confined to a thin film along the parting planes." In these shales, cleavage planes are more closely spaced in belts of intense folding, and, under the same structural conditions, they are more closely spaced in more argillaceous beds than in more calcareous beds. Fracture cleavage plates in the argillaceous limestone of the Stony Point formation commonly range from one half inch to 5 inches thick. Fracture cleavage in calcareous shale is finer, and in the noncalcareous shale of the Iberville the planes are so close as to resemble flow cleavage (Swanson, 1941, p. 1246), but in thin section cut perpendicular to the finest cleavage, it is seen to be composed of somewhat irregular and discontinuous joint-like fractures 0.02 to 0.06 mm. apart. Bedding displacements of 0.01 to 0.04 mm. occur along the cleavage planes (Hawley, 1957, p. 82).

Although innumerable faults cut the shales, only a few displace them enough to juxtapose different formations. On most faults the rock of both walls is so similar that only minor displacement can be assumed. Block faulting typical of the western and southern Champlain Valley is distinct only in the older Trenton, Chazy, and Canadian formations of western South Hero, where Kay and his former students have mapped them (personal communication). Shear along bedding surfaces, cleavage surfaces, and at varying angles to both is very common. In more intensely folded belts, multiple shears occur along crests and troughs of folds. The bearing of slickensides is remarkably constant, regardless of the attitude or type of surface on which movement occurred. Of 119 slickensides bearings measured in this area, only three lay outside the arc between N 25° W and N 85° W (Hawley, 1957, p. 81).

Field Trip Stops

The best exposures of the shales and limestone are along the lakeshore bluffs. During the spring months and after long periods of heavy rain, the lake may be higher than normal, and many of these exposures may be inaccessible. Field localities are shown in Fig. 1. THE STOPS ARE ON PRIVATE LAND. PERMISSION HAS BEEN OBTAINED FOR THE STOPS WE WILL

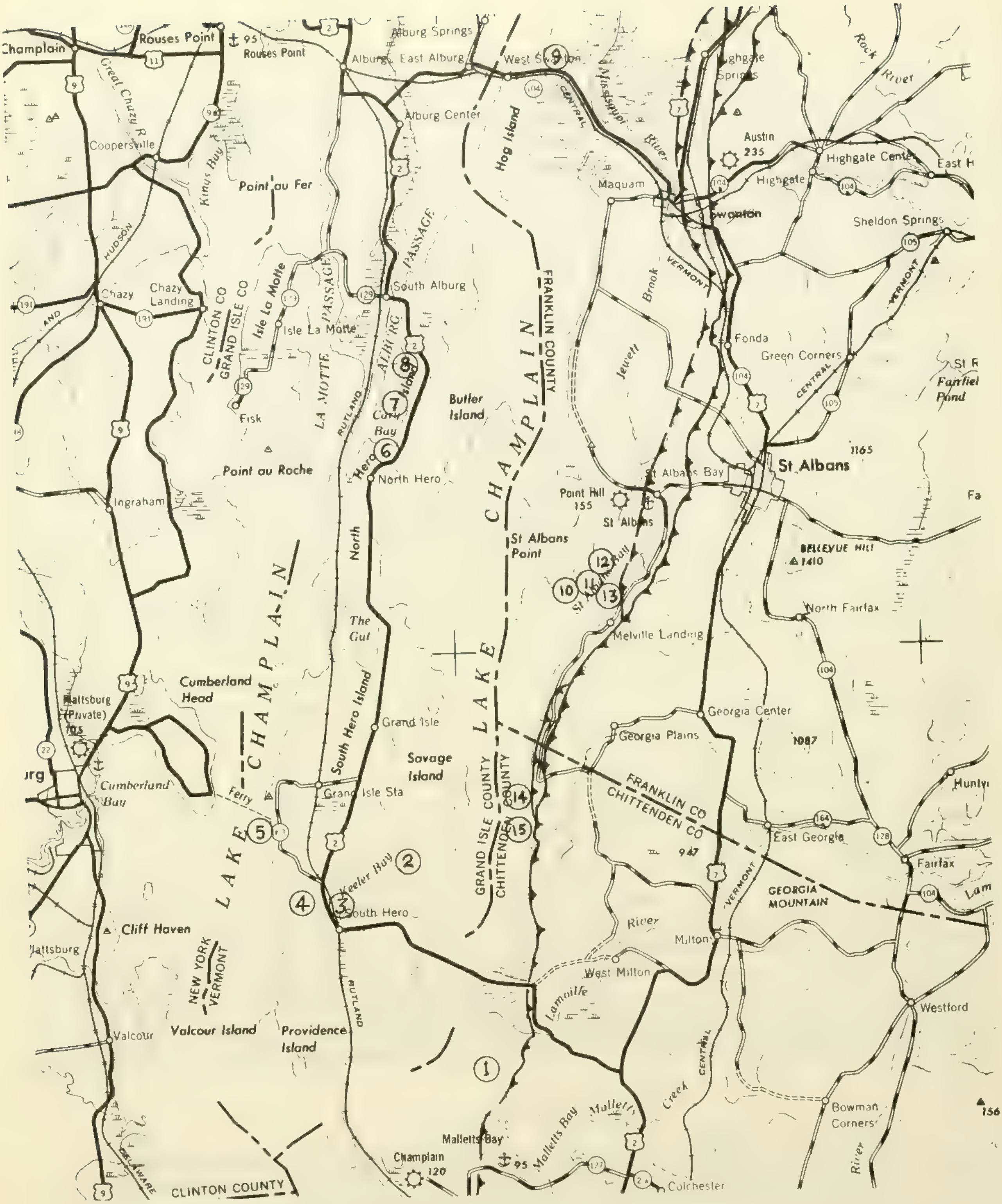

Figure 1. Trip 5 - Field trip stops.
Scale: 1 in. = 4 mi.

VISIT. THOSE WHO MAY WISH TO VISIT THESE LOCALITIES IN THE FUTURE SHOULD GAIN PERMISSION FOR EACH VISIT. GIVE GEOLOGY A GOOD NAME BY BEING VERY THOUGHTFUL.

Figure 2. Stop 1, Clay Point. Scale: 1:24,000.

Stop 1. Clay Point, between Malletts Bay and the Lamoille River, east shore of lake. (Fort Ethan Allen Quad., 1:24,000). THIS PROPERTY IS POSTED, AND PERMISSION MUST BE OBTAINED. In the transition beds in the lower Iberville (interbedded calcareous and noncalcareous shale, with argillaceous limestone, argillaceous dolomite, fine-grained dolomite, and silty-laminated dolomite with current cross-bedding) there is a small, overturned anticline cut by small thrust faults. The relationship of cleavage to bedding, plunge of the fold, identification of tops by cross-bedding, and the faulting make this a worthwhile stop for a structural geology class.

Stop 2. From Kibbee Point (northeastern South Hero) southeastward along the shore for 2500 feet, is exposed the transition from Stony Point to Iberville formations. (South Hero Quad., 1:24,000). With a few minor rumples the dip is southeastward all the way to a deep gully and small bay which separate a steep bluff-point to the east from the shore northwestward to Kibbee Point. This bluff, 2800 feet SE of Kibbee Point is composed of Stony Point argillaceous limestone and calcareous shale, overturned and dipping 55° southeastward. Thus, the gully conceals the faulted core of an overturned syncline. The fault is very likely a thrust, east side up.

West of the gully is Iberville, about 90% finely cleaved noncalcareous shale, with interbedded silty cross-laminated dolomite. Northwest from here to Kibbee Point the proportion of calcareous shale

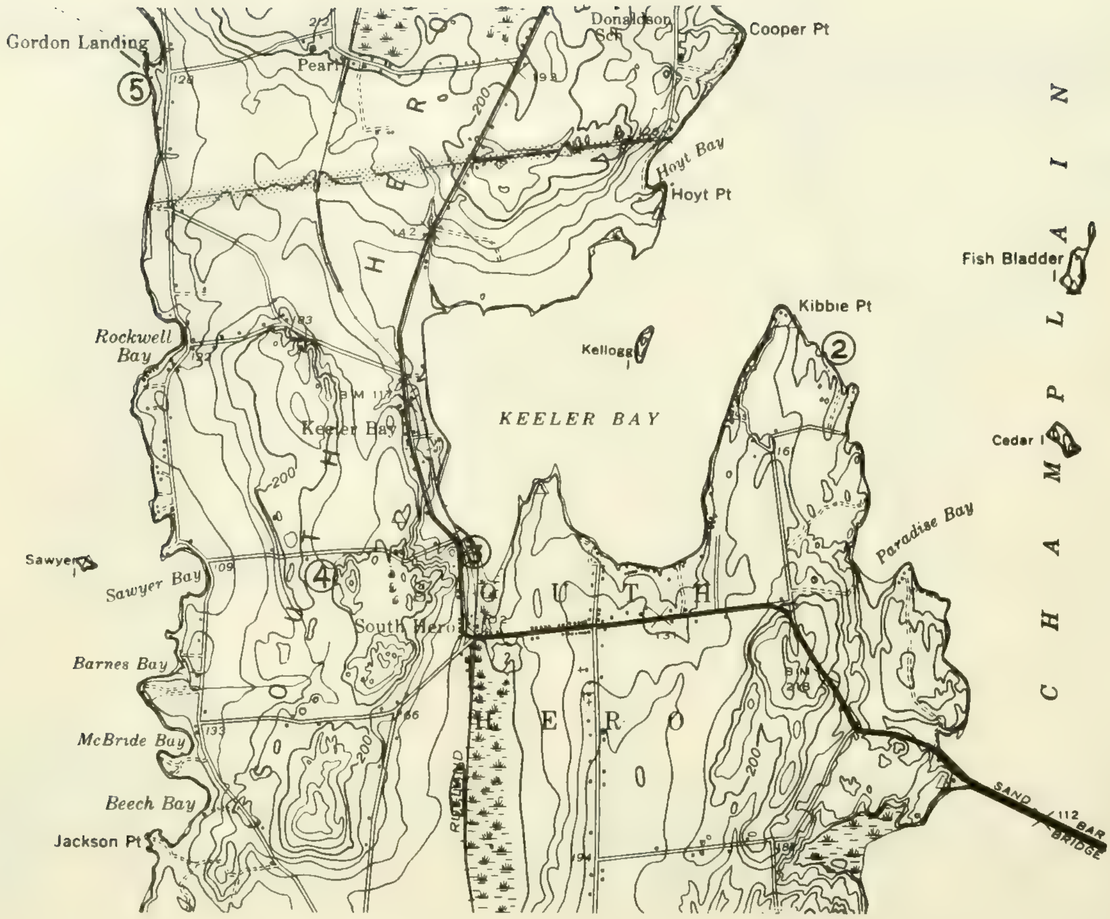

Figure 3. Stops 2-5, South Hero Island. Scale: 1:62,500.

increases. About 220 feet southeast of Kibbee Point the southeastward-dipping beds are massive calcareous shale (Stony Point fm.). About 900 feet south of Kibbee Point on its west shore the Stony Point beds still lower in the section are predominantly argillaceous limestone, interbedded with calcareous shale.

Stop 3. Road cut on US 2, at the southwest corner of Keeler Bay, 1 mile south of junction with Sunset View Rd. (South Hero Quad., 1:24,000). The road cuts northward across a NE'ward plunging anticline in the Stony Point formation. The rock is dominantly silty-laminated argillaceous limestone with some calcareous shale, and is inferred to be in the thick argillaceous limestone zone in the middle of the formation. There are conspicuous

bedding-plane slickensides on the west side of the road. Rotational offset along fracture of cleavage can be seen by matching silty laminae across the fractures. On the east side of the road, harder fine-grained limestone beds (5" ±) have buckled and overlapped.

Stop 4. Small quarry in Glens Falls Ls., .1 mile S of Sunset View Road, .6 west of US 2. (South Hero Quad., 1:24,000). This thick-bedded limestone with fossiliferous zones (1-3") at intervals of 1 to 5 inches, will serve to dramatize the change to predominantly shaley rocks in formations younger than the Glens Falls. The area of the quarry has been mapped as the Larrabee member. (Erwin, 1957).

Stop 5. West shore of South Hero Island, extending for one mile southward from the breakwater at Gordon Landing. (South Hero Quad., 1:24,000). The lower 215 feet of the Stony Point formation is exposed between the breakwater and the top of the Cumberland Head formation, 2900 feet to the south. In the next 2300 feet of shoreline, the upper 145 feet of the Cumberland Head formation is exposed. These sections are described in the text article. The south end of this section is cut off by a right lateral wrench fault striking N 59° W, dipping 79° NE. South of the fault the interbedded limestone and shale (about 79% ls., 30% sh.) have been mapped as the Shoreham member of the Glens Falls formation (Erwin, 1957) on the basis of lithology and the presence of Cryptolithus.

Stop 6. Road cut on east side of US 2 halfway between City Bay (North Hero Beach roadside park) and Carrying Place. (North Hero Quad., 1:24,000). This outcrop shows the interbedded laminated argillaceous limestone and calcareous shale typical of the middle section of the Stony Point formation. It lies close to the axis of a major, northeastward plunging anticline.

Stop 7. Middle point on north side of Cary Bay, North Hero, 2000 feet east of Blockhouse Point. (North Hero Quad., 1:24,000). Typical Iberville cyclic bedding is exposed for about 1500 feet along this shore, extending eastward from the place where the access road meets the shore. From west to east are: an asymmetrical syncline, an asymmetrical anticline, and to the east of a covered interval is the east, overturned limb of a large syncline. These folds are in the axial area of a large, northeastward plunging, overturned syncline. Relationships of cleavage to bedding, axial surfaces, and direction of plunge are well shown. Small-scale current cross-lamination on some beds indicates southwestward flow.

Stop 8. Quarry in Iberville (mislabelled "gravel pit" on No. Hero Quad., 1:24,000), 1.6 miles S 10° E from east end of North Hero-Alburg bridge. The beds are almost flat-lying, and only about 15 feet of section is exposed, but it is typical cyclic deposition, and the details are well shown.

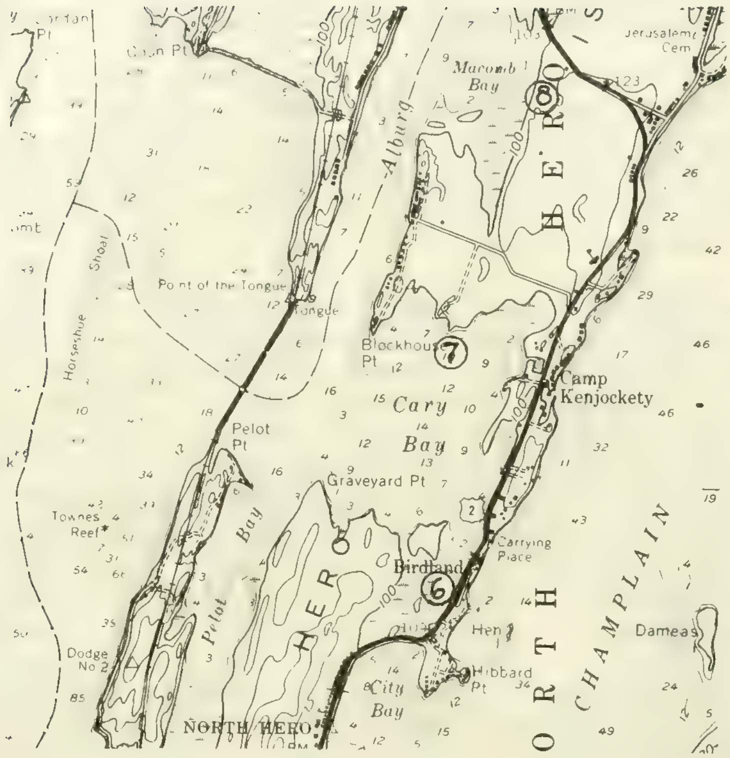

Figure 4. Stops 6-8, North Hero Island. Scale: 1:62,500.

Stop 9. Upper Iberville beds in quarry (mislabelled "sand and gravel pit" on East Alburg Quad., 1:24,000) 1800 feet north of Vt. Hwy 78 and 600 feet west of Campbell Road, northern Hog Island, West Swanton. The quarry exposes an overturned anticline, thrust faulted on the upper, eastern limb, with adjacent syncline immediately westward, also faulted.

Stop 10. Southernmost tip of St. Albans Point, on property of former Camp Kill Kare, now a state park. (St. Albans Bay Quad., 1:24,000). Northeastward plunging asymmetrical anticline with linked small syncline northwest of it, in Iberville noncalcareous and calcareous shale with dolomitic interbeds.

Stop 11. Between Camp Kill Kare's access road and the lake, about halfway between the private cottages and the Camp buildings. (St. Albans Bay Quad., 1:24,000). There are 31 feet of white weathering, grayish-black chert in beds of 2 to 6 inches, dipping steeply (69°) southeastward on the southeast flank of the anticline at Stop 9. Structurally overlying the chert beds is

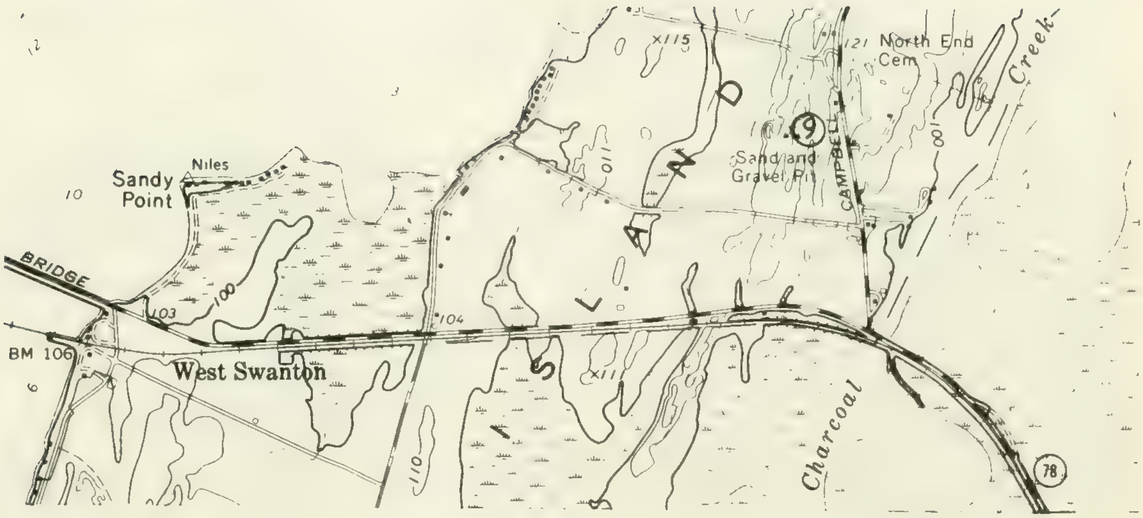

Figure 5. Stop 9, Northern Hog Island. Scale: 1:24,000.

Figure 6. Stops 10-13, St. Albans Bay area. Scale: 1:24,000.

black siliceous argillite in which bedding is not apparent because of its irregular, chippy foliation. The argillite contains rounded pebbles (avg. 1 by 2 inches) of gray dolomite and fragments of chert. Some graptolites were found in the argillite, but smearing precluded identification. This is part of the Hathaway formation. It is likely that the chert beds here represent a larger mass involved in a submarine slide.

Stop 12. Hathaway Point, at the south end of St. Albans Point. (St. Albans Bay Quad., 1:24,000). This is the type locality for the Hathaway formation. It has a matrix of pale-greenish-yellow weathering rock seen on a polished surface to be composed of small, irregular, curdled masses of greenish-gray to olive-gray argillite. Streamed and isoclinally folded in the matrix is black siliceous argillite similar to that associated with the chert beds at Stop 11. "Floating" in the matrix are small masses of grayish-black radiolarian chert which are commonly angular, as well as masses of bedded chert measurable in tens of feet. Fragments of dolomite and dolomitic siltstone occur in the western part of the Hathaway point exposure. Numerous slickensided tectonic shears are present in a variety of orientations. One 40-foot wedge between shears is composed of isoclinally folded calcareous and noncalcareous shale with occasional boudinaged masses of fine-grained limestone, resembling the transition beds at the base of the Iberville. Both of the islands east of Hathaway Point, in the middle of the bay, are composed of chaotically deformed argillite and chert. It is assumed that St. Albans Bay may lie over a deep synclinorium.

Stop 13. Lime Rock Point, on the southeast side of St. Albans Bay. (St. Albans Bay Quad., 1:24,000). At the base of the bluff composed of the Beldens (Upper Canadian) crystalline limestone with buff-weathering dolomitic beds, there is a dramatic exposure of the Highgate Springs overthrust; lower Ordovician Beldens Limestone over upper Ordovician Iberville calcareous and noncalcareous shale with occasional beds of yellowish-brown weathering fine-grained dolomite and silty dolomite. At the base of the high, steep bluff about one half mile to the east is the Champlain overthrust, on which the lower Cambrian Dunham dolomite is thrust westward over the Beldens. South of Lime Rock Point the Highgate Springs thrust slice is overlapped by the Champlain thrust for two and a half miles. It reappears for four miles, and then disappears again under the Champlain thrust, southeast of Beans Point. This is as far south as the Highgate Springs slice can be traced.

Stop 14. Beans Point, east shore of lake in northwest Milton. (Georgia Plains Quad., 1:24,000). The Hathaway crops out intermittently for 1200 feet north from Beans Point. This is in a zone of intense deformation close to the Highgate Springs thrust, the trace of which is covered, probably about 600 feet back from the shore. The base of the steep bluffs 2000 feet back from the shore marks the trace of the Champlain fault, on which lower Cambrian Dunham dolomite has been thrust over Beldens crystalline

Figure 7. Stops 14 & 15, Northwestern Milton. Scale: 1:62,500.

limestone and dolomite of the Highgate Springs slice.

The Hathaway is composed of boulders and fragments "floating" in mashed argillite. The argillite is mottled olive gray to dark greenish gray to greenish black. On a polished surface cut perpendicular to foliation the mottled colors are seen to represent original bedding which has been folded most intricately, and sheared with no development of slickensides or breccia. The small-scale shearing has completely healed, and some minute fold crests merge into the adjacent bed, a streaming of one bed into the next with no sharp boundary. Included in the argillite are rounded fragments of moderate-yellowish-brown weathering, dark gray fine-grained dolomite and cross-laminated dolomitic siltstone, sub-angular to rounded, up to 4 by 7 by 20 inches in size. The long axes of the boulders are approximately parallel, plunging about 55° toward S 45° E. Foliation causes the argillite to split into irregular tapered chips. Thirty-six feet of cover separates the north end of the Hathaway outcrop from cyclic-bedded upper Iberville which lies overturned, dipping 46° northeastward.

<u>Stop 15.</u> Camp Watson Point, 3/4 mile south of Beans Point (Stop 14). (Georgia Plains Quad., 1:24,000). The core of a large, overturned syncline is exposed on the point, plunging 18° toward N 56° E. The overturned limb, dipping 29° southeastward, is exposed for 200 feet or more along the shore to the south. The rock is lower Iberville transition, with interbedded calcareous and noncalcareous shale, argillaceous limestone, and silty laminated dolomite.

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Trip 7

ROTATED GARNETS AND TECTONISM IN SOUTHEAST VERMONT

by

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Conventional structural and stratigraphic data (fig. 1--based primarily on Doll et al., 1961) and their topological implications suggest that during the late Paleozoic two large, recumbent, isoclinal, sigmoid folds in Paleozoic, metamorphosed, stratified rocks of southeast Vermont (Table 1) predated the mantled gneiss domes with which they are associated. One of these folds involved units stratigraphically and structurally beneath the Siluro-Devonian calcareous and non-calcareous schists of the Waits River formation. The other, involving Siluro-Devonian units structurally above but of otherwise undemonstrated stratigraphic relationship to the same unit, is exposed in culminations associated with seven domes. The approximate axial parallelism of these folds and the opposite rotations of their short limbs suggest that these folds resulted from westward extrusion of rocks in between.

Using methods described elsewhere (Rosenfeld, 1970), regional study of spirally arranged inclusions in garnets and rotations determined therefrom confirms the early presence within the Waits River formation of a surface, quasi-parallel to the bounding strata now exposed to the east and west, across which the rotational senses possessed mirror symmetry (Rosenfeld, 1968). After graphical correction for effects resulting from rise of the gneiss domes, the rotational axes of the garnets parallel those of the giant recumbent folds; and the rotational senses of the garnets are those to be expected from flexure slip folds having the observed rotations of their short limbs (fig. 2; Rosenfeld, 1968, p. 193).

In further accord with the westward extrusion are:

(1) pebbles and former phenocrysts in the eastern part of the area extremely elongated in the direction of extrusion; also a prominent mineral lineation in the same direction (observable on Putney Mountain-Windmill Mountain Ridge; Rosenfeld, 1968, p. 197-199);

(2) boudinage in the eastern part of the area with fractures quasi-parallel to the corrected rotational axes of the garnets; these give way to the west to compressional folding of the same orientation (observable in central Vermont just east of the Green Mountains);

(3) an angular divergence in a westerly direction of about 10 degrees between the Standing Pond formation on one side of the Waits

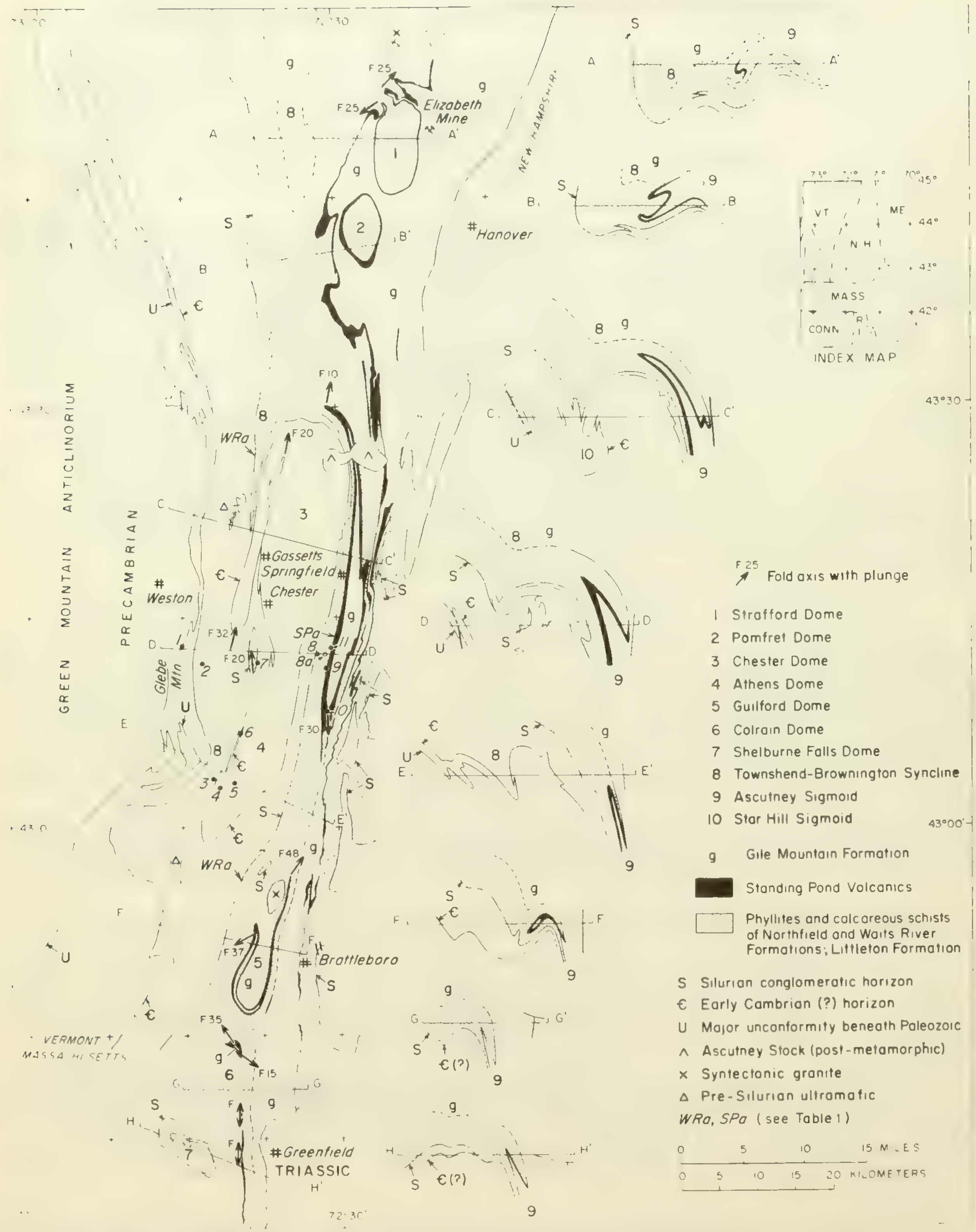

Fig. 1 - Field trip stops italicized

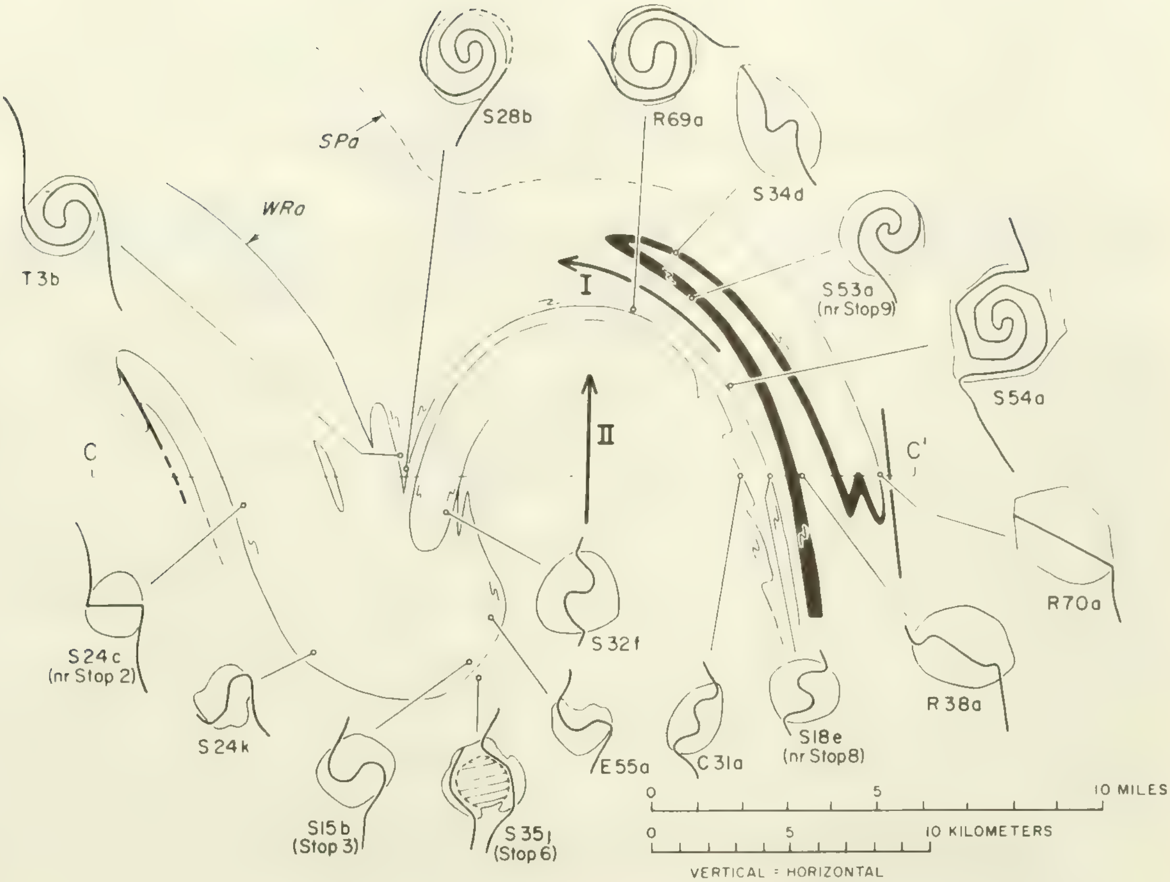
SPa
S28b
R69a
S34d
WRa
T3b
I
S53a
(nr Stop9)
S54a
II
C
C'
S24c
(nr Stop 2)
S32f
R70a
S24k
E55a
C31a
S18e
(nr Stop8)
R38a
S15b
(Stop 3)
S35j
(Stop 6)
0 5 10 MILES
0 5 10 KILOMETERS
VERTICAL = HORIZONTAL

Fig. 2

Table 1

Condensed Chronologic Table of Metamorphosed Rocks
(See Doll *et al.*, 1961, for more details)

Geologic Age	Unit or Feature	Lithology
Devonian (?)	New Hampshire Plutonic Series	Late synkinematic granitic rocks
Devonian*	Gile Mountain Formation, Littleton Formation (?)*	Quartzo-feldspathic schist, graphitic schist, some calcareous
Siluro-Devonian	Standing Pond Volcanics	Chiefly amphibolites, greenschists of volcanic origin
	UNCONFORMITY (?) (*SPa*)	
Siluro-Devonian	Northfield and Waits River Formations, Littleton Formation (?)* (*WRa*)	Graphitic calcareous and non-calcareous schist
Silurian	Shaw Mountain Formation	Quartz conglomerate, porphyritic volcanics
	UNCONFORMITY	
Late Ordovician	Ultramafic intrusives	Dunite, serpentinite, steatite
Early Cambrian to Mid-Ordovician	Pinney Hollow through Missisquoi Formations	Heterogeneous schists, hornblende gneisses and amphibolites
Late Precambrian to Early Cambrian	Cavendish through Hoosac Formations	Augen gneiss, conglomerate gneiss, albitic and paragonitic schist, dolomite
	MAJOR UNCONFORMITY	
Precambrian	Mt. Holly complex	Assorted gneisses, granites, schists, amphibolites, and marbles

* The direction of facing across the Standing Pond Volcanics is still uncertain. I have followed Chang *et al.*, 1965, in elevating this unit to formational status. Implications of the alternative possibilities are discussed in their paper (p. 40, 56-62).

River formation and the Shaw Mountain formation stratigraphically beneath it (further possible indication of this divergence appears in the easterly offset of negative gravity anomalies (Bean, 1953, p. 528-533) in the Strafford and Pomfret domes). This divergence is particularly evident south of the Ascutney stock (fig. 1);

(4) "downstream" folding oriented in a westerly direction to the west of a large pre-Silurian dunite mass (observable in the eastern part of the Wilmington Quadrangle west of the East Dover ultramafic body mapped by Skehan, 1961).

Analysis of the rotations represented by the garnets indicates that the Green Mountain anticlinorium, although present in older stratigraphic units at the time, manifested itself as a rejuvenated anticlinorium within the Siluro-Devonian strata only after the westward extrusion and contemporaneously with the development of the mantled gneiss domes to the east. The anticlinorium therefore did not form a barrier to the westerly extrusion and consequent loading of areas to the west. This earlier anticlinorium may be related to an earlier Paleozoic metamorphism evident in rotated garnets containing growth-rotation "angular unconformities" (loc. S35j, fig. 2).

The shear senses and orientations of conspicuous minor folds, commonly at high angles to the early (inner) rotational axes of the garnets, give evidence of later up-thrust of the gneiss domes (fig. 2; Rosenfeld, 1968, p. 193). The surfaces included in the outer parts of the garnets also reflect gradual transition to late rotations of the garnets about axes parallel to the folds and of similar rotational senses. The high angle between the early and late rotational axes of the garnets, both of which must have paralleled the schistosity at their respective times of growth, permits apportionment of the rotation. On the east limb of the Chester Dome, garnets at one locality show 625° rotation for the early stage of deformation and 105° for the late stage.

Interpretation of the proximate mechanism of diastrophism for the early and major diastrophic event depends primarily upon knowledge of the as yet unknown age relationship of the units bounding the Waits River formation on the east. If these units should prove older than the Waits River formation, the indicated westward transport of material may be ascribed to flexure-slip folding of the westward-opening lower half of a giant, initially recumbent, sigmoid fold whose upper half is nowhere exposed in eastern Vermont. If the same units should prove younger than the Waits River formation, the transport may be ascribed to westward intrastratal extrusion of the relatively dense Waits River formation, possibly down a gently inclined slope tilted toward the west. It is thus of great importance to resolve this ambiguity by development of procedures for resolving the above stratigraphic uncertainty.

Road Log for Trip 7

Road log begins on Route 11, just west of the summit of the pass over the east range of the Green Mountains, 0.5 miles southwest of North Windham, Vermont, about 200 feet west of the west boundary of the Saxtons River Quadrangle. This excursion is basically a "no hammer" trip, although collecting at Stop 3 is all right. I'd like to enlist the assistance of all participants in helping to preserve the highly visible minor structural features so that future geologists will be able to see them in their field context. May these features avoid the "tragedy of the commons!" Names of units and major structural features referred to below are largely from Doll et al., 1961, and Rosenfeld, 1968. A perusal of these references before undertaking this excursion will be helpful.

Mileage

0.0 STOP 1. Angular unconformity between the overlying prograde metamorphosed conglomerates of the Tyson formation and the underlying retrograde metamorphic rocks and pegmatites of the Precambrian Mount Holly complex. This unconformity is significant for this trip because it demonstrates the direction of stratigraphic "tops." Proceed easterly on Route 11 through the Hoosac formation.

0.6 North Windham. Turn right onto Rt. 121.

0.8 Northernmost exposures of Turkey Mountain member (amphibolite) of Hoosac formation outcrop in draw to west. Continue through schists of Pinney Hollow formation.

2.2 Near crest are exposures of Chester amphibolite member of Pinney Hollow formation. Strong down-dip lineation of pale green amphiboles. Continue through Ottauquechee and Stowe formations.

2.6 STOP 2. First rotated garnet locality in outcrop at northwest corner of intersection. Garnets in schist of Stowe formation show small counterclockwise rotation after growth about nearly horizontal axes when viewed in a northerly direction (the direction of view used subsequently unless otherwise stated). Proceed southerly from Rt. 121 on Windham Road past outcrops of Stowe formation and rusty shales of the Whetstone Hill member of the Missisquoi formation.

4.0 Windham Center. From here almost to South Windham, the road lies within the banded rusty-weathering graphitic schists of the Ottauquechee formation. Rise in metamorphic grade is most evident in the field in the transition from pale green amphibolites to dark green to black amphibolites. Near Windham Center we cross the oligoclase isograd, northwest of which plagioclase more calcic than nearly pure albite is not found, regardless of bulk composition of the rock. This isograd is related to a miscibility gap within the plagioclase feldspar series.

7.6 South Windham. Chester amphibolite.

8.0 Jamaica-Townshend town line. Enter the typical green garnet-magnetite-chlorite-sericite schist comprising the main part

of the Pinney Hollow formation and through which the road passes for the next 2.0 miles.

10.0 Turkey Mountain member appears on ridge to west. From here to West Townshend we pass from the Pinney Hollow formation into the characteristic albite schists of the Hoosac formation.

10.7 West Townshend, ancestral home of the Tafts of Ohio. Turn left onto Rt. 30 and proceed southerly through a tectonically compressed section from Hoosac to the base of the Missisquoi formation.

11.1 Base of Missisquoi formation. Continue in typical "pinstripe" quartzofeldspathic schists of Moretown member of Missisquoi formation.

11.6 Roadcuts on west side of highway show eastward dipping beds of "pinstripe" in Moretown with a prominent boudinage fracture of horizontal orientation. Continue in highly contorted schists and amphibolites of the Moretown across the axis of the Townshend-Brownington syncline onto the west limb of the Athens (pronounced Aythens) dome.

12.8 Thin amphibolites in smooth outcrops of Moretown on the left exhibit boudinage.

13.1 Park cars in parking area on right at Townshend Flood Control Dam. STOP 3 is in the roadcut on the northeast side of the highway opposite the dam. Rotated garnets showing counterclockwise rotation on the west limb of the Athens dome. Note the relative consistency of the shear sense indicated by the rotated garnets in contrast to that of the drag folds. The origin of this contrast has been discussed elsewhere (Rosenfeld, 1970, p. 92). Garnets observed here are believed to have grown and rotated before development of the Athens dome during the lateral extrusion toward the west. The relict "oligoclase isograd" may be observed in the form of coexistent albite, oligoclase, and clinozoisite encapsulated in garnets at this locality (Rosenfeld, 1970, p. 90-91), even though the staurolite isograd is only a few tens of feet to the east. Note the large boudinage fractures in amphibolites here. Proceed southeasterly on Rt. 30 through a compressed but apparently complete section from the Moretown to the Hoosac formation.

13.5 Scott Covered Bridge on right. Amphibolite in what is believed to be Hoosac formation on left. If these rocks correlate with the main band of the Hoosac to the west, they are of a distinctly more banded and gneissic facies. Just beyond the bridge on the left are some very nice secondary drag folds on a large fold, incompletely exposed in the outcrop.

13.8 STOP 4. Conglomerate gneiss of Tyson formation (?) on west in contact with Bull Hill gneiss member of Cavendish formation. The Bull Hill gneiss characteristically has coarse microcline augen and is of granitic composition. However, it is also a rather widespread stratigraphic unit on the

Chester and Athens domes. It is therefore possible that the Bull Hill gneiss represents a metamorphosed stack of rhyolitic volcanics. In the southern part of the Athens dome, it has not been possible to delineate accurately the boundary between the Bull Hill gneiss and what are believed to be older but lithologically similar Precambrian granitic augen and flaser gneisses in the core of the dome. Note counterclockwise drag folds in gneiss, believed to be a result of upthrusting of the gneissic core of the dome. Proceed easterly on Rt. 30 through broad zone of granitic gneisses to

15.0 Townshend. Turn left off Rt. 30 onto Rt. 35 and proceed northerly.

15.4 STOP 5. Outcrops lie across the field to the west and consist of magnetite-bearing granite flaser gneiss, believed to have been the relatively low-density "plunger" accounting for the buoyant upward thrust of the Athens dome. Continue north on Rt. 35 through heterogeneous gneisses, some rusty weathering and containing coarse graphite flakes rather like the Washington gneiss described by Emerson in the Berkshires.

16.9 Simpsonville.

18.4 Easy to miss intersection. Bear left off Rt. 35 onto Grafton Road.

18.6 For the next 0.2 miles, passing through a band of calc-silicate rocks, characterized by coarse graphite flakes and pyrrhotite, that strikes northeasterly through the core gneisses of the Athens dome at a large angle to the mantling strata. This discordance provides, perhaps, the best evidence to date that the core gneisses of the Athens dome lie unconformably beneath the mantling strata.

18.8 Continue through banded, contorted, biotite gneisses of the core of the Athens dome.

19.6 Top of grade. Bull Hill gneiss on dip slopes along east side of South Branch of Saxtons River to north. Valley probably owes its alignment to an easily eroded dolomite (observable at a number of localities on Rt. 35 north of Grafton) that separates albite schist of the Hoosac formation on the west from the Bull Hill gneiss.

20.3 Easy to miss turn. Turn sharply left onto single lane, steep dirt road (Acton Hill Road). Proceed through Hoosac formation.

20.6 Cross brook.

20.9 East contact of garnet-kyanite- staurolite-paragonite schist of Pinney Hollow formation in core of anticlinal portion of Ober Hill fold. Pass across Ober Hill fold.

21.6 Intersection. Let lead car turn around before entering intersection. Then, one by one, each car should turn left, then back up sufficiently far to make room for following cars to do same. Continue back down the Acton Hill Road, following lead car.

21.8 Park your car as far off the road to the right as possible. STOP 6, exhibiting garnets with angular growth unconformities, is on the ledges visible to the southwest of the road (Rosenfeld, 1968, p. 196). The rock is a garnet-staurolite-paragonite-muscovite schist. Chloritoid and staurolite exist as an armored relict assemblage in the garnet. There is no chloritoid outside the garnet. The earlier garnet probably grew during the Taconic orogeny or possibly during an earlier orogeny. Proceed back toward Townshend-Grafton Road.

22.1 On the left are some remarkably fine counterclockwise drag folds, some of which have transcurrent "slip fractures" of similar shear sense about the same axis. These fractures provide evidence of the "lateness" of these folds.

22.9 Townshend-Grafton Road. Turn left and continue north.

27.7 Grafton, a picturesque village in which some of the finer examples of old Yankee architecture have been restored and preserved by the liberal application of dollars. Turn left onto Rt. 121, passing successively through a rather complete section of units from the Hoosac formation to the rusty-weathering, graphitic schists of the middle Ordovician Cram Hill member of the Missisquoi formation.

29.9 STOP 7. Westward dipping beds of conglomeratic quartzite and interbedded garnet-muscovite schist of Silurian Shaw Mountain formation. These beds lie on the east limb of a syncline (Spring Hill syncline) whose axial surface dips to the west. This syncline is believed to be the detached (by megaboudinage) westward-opening, lower part of the Star Hill sigmoid (Figure 1, Section D-D'). It contains in its core a section of dense amphibolites that is thicker than usual within the Shaw Mountain formation. This dense mass, in the "keel" of the formerly westward opening fold, is believed to have "hinged" downward clockwise during the doming stage. Thus, the exposure at this stop is believed to be a relict of the short limb of the Star Hill sigmoid. In support of this interpretation is the sequence of rotations found in garnets within a schistose parting of Shaw Mountain quartzite a mile to the north--early clockwise, late counterclockwise. "Unrotating" the late rotation at this exposure aligns the elongate pebbles in a west-southwest orientation, the direction of lateral extrusion. Some of the quartzite at this locality contains coexistent staurolite and chloritoid, a not rare assemblage in this unit. Turn around and return to

32.1 Grafton on Rt. 121, continuing through the village across the Saxtons River and turning left onto

32.2 Rt. 35, proceeding northerly along approximately the same stratigraphic horizon that was followed south of Grafton. Bull Hill gneiss to east.

33.9 Dolomite under albite schist of Hoosac formation on left. Leaving Athens dome; entering Chester dome.

36.4 Enter Grafton Gulf.
36.9 Leave Grafton, Windham County; enter Chester, Windsor County.
37.0 Note pillar of dolomite supporting albite schist on left, dip slope of Bull Hill augen gneiss on right.
37.5 Summit of Grafton Gulf.
38.3 Leave Saxtons River Quadrangle; enter Ludlow Quadrangle.
39.5 Chester. Turn right onto Rt. 103.
40.9 Return to Saxtons River Quadrangle.
42.3 Bull Hill gneiss on east limb of Chester dome.
42.5 Enter town of Rockingham, Windham County. Crossing Hoosac formation.
42.7 Crossing from Pinney Hollow through intermediate units into Missisquoi formation.
44.3 Easy to miss intersection. Turn sharp left off Rt. 103 onto dirt road with bridge over railroad tracks.
44.4 "Vermont Beautiful" on left!
44.5 Covered bridge.
44.8 Crossing Shaw Mountain formation--not exposed near road.
44.9 STOP 8. Ledges in woods north of road. Sieve texture garnets in calcareous schists of lower Waits River formation showing early counterclockwise rotation (Event I; conspicuous) followed by late clockwise rotation (Event II; observed with difficulty). Continue easterly.
45.7 Optional STOP 8a. Main zone of calcareous schists with subordinate phyllites within Waits River formation. One of the best exposures of the Waits River formation in southern Vermont. Big sprays of zoisite. Isoclinal folding. Easily observed rotated garnets. Mafic dike with calcite phenocrysts. Turn right across bridge and railroad tracks.
46.0 Turn left onto Rt. 103.
46.2 Turn right off Rt. 103 onto Pleasant Valley Road. Passing through heterogeneous rock types of Standing Pond formation, mostly mafic volcanics.
47.1 Turn right off the Pleasant Valley Road onto single lane dirt road.
47.2 Park cars and proceed northerly across field about 1,500 feet into woods just northwest of northwest corner of field to STOP 9 at contact between garnetiferous phyllite of Waits River formation on west and coarse garnetiferous schist of the Standing Pond formation containing sprays of hornblende (fasciculitic schist or "garbenschiefer"). Large garnets show a single large clockwise rotation associated with Event I, in contrast to those at Stop 8. A photograph of a rotated garnet from this locality appears as figure 14-6 in Rosenfeld, 1968 (p. 195). Evidence of Event II at this locality appears only as gently northward plunging crinkles. For further discussion of this locality, see Rosenfeld, 1970, p. 89. Return to Pleasant Valley Road by car.
47.3 Turn right onto Pleasant Valley Road.
48.7 Septum of Waits River-like calcareous schist and phyllite in Standing Pond formation.

48.8 Exposures of banded and massive amphibolites of Standing Pond formation near eastern contact with Gile Mountain formation. Clockwise drag folds. Road continues southerly along east side of Standing Pond formation.

51.0 Intersection with Rt. 121. Continue east on Rt. 121.

51.3 Village of Saxtons River. Park cars. STOP 10. The purpose of this stop is to observe southward plunging minor folds in the Standing Pond formation along the axis of the upward closing fold (anticline) of the Ascutney sigmoid. The axis at this horizon reappears to the south on the Guilford dome near the syntectonic Black Mountain granite in Dummerston (fig. 1). Folds with counter-rotating garnets on their limbs appear along the north side of the river, 0.3 miles to the west (Rosenfeld, 1970, p. 85-86). Turn westerly on Rt. 121.

51.6 Bear right off Rt. 121 onto Pleasant Valley Road.

56.2 Turn left off the Pleasant Valley Road onto Rt. 103.

56.3 Turn right off Rt. 103 toward Brockways Mills, continuing across bridge past Stop 8a and to the right on paved road toward Springfield.

57.5 Park. Proceed westerly across north end of field past small cottage to STOP 11 at contact between garnetiferous phyllite of Waits River formation and "garbenschiefer" with large garnets. This locality is, perhaps, the best locality for seeing evidence of both Events I and II within a single rotated garnet. A stereoscopic photograph of a rotated garnet from this locality appears as figure 14-3 in Rosenfeld, 1968, p. 192. A discussion of the generation of the central surface of garnets at this locality is found in Rosenfeld, 1970, p. 40. The trip ends at this locality.

To get to Burlington, about 120 miles away, return to Rt. 103, turn left, and get onto Interstate 91 North at Interchange 6. Turn left onto Interstate 89 at White River Junction. Interstate 89 will take you to Burlington.

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Trip B-8

STRATIGRAPHIC AND STRUCTURAL RELATIONSHIPS ACROSS THE GREEN MOUNTAIN ANTICLINORIUM IN NORTHCENTRAL VERMONT

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Introduction

This road log provides a guide for a field trip which extends from East Georgia to Hardwick, Vermont, along the Lamoille River. The stops were chosen to provide some understanding of the stratigraphic and structural relationships of the Cambrian and Ordovician rocks on both flanks of the Green Mountain anticlinorium and of the problems involved in the correlation of these rocks across the crest of the anticlinorium.

Geologic mapping in this area preceded the compilation of the Geologic Map of Vermont (1961); the sketch map indicates only the route and major stops, and the State Geologic Map will serve as the basic map reference for this trip. The geologic reports on the four quadrangles involved have all been published: Milton quadrangle (Stone and Dennis, 1964); Mt. Mansfield quadrangle (Christman, 1959); Hyde Park quadrangle (Albee, 1957); Hardwick quadrangle (Konig and Dennis, 1964). In addition, Osberg (1969) gives a concise summary of the geology of this area and a reinterpretation of the patterns on the State Geologic Map. Albee (1968) describes the metamorphic zoning in northern Vermont. Since these reports are readily available, this road log will contain few details but will emphasize the need for additional detailed mapping in several areas to solve certain critical correlation and structural problems.

The Green Mountain anticlinorium extends the full length of Vermont. Sequences on both its eastern and western flanks are rather uniform so that individual formations can be traced from Massachusetts, through Vermont, and some distance into Quebec. Rapid east-west facies changes, extensive unconformities, and thrust faults of unknown extent have been utilized in the correlation of these two sequences, but there is no generally accepted detailed correlation of the lower Paleozoic (pre-Shaw Mountain) units.

In the absence of fossils, correlation can be attempted only by relating detailed lithologic characteristics and sequences or by tracing units along strike to points where they "bridge" the Green Mountain anticlinorium in an axial depression. The correlation on

the State Geologic Map of the Ottauquechee Formation with the Sweetsburg Formation and its upper Cambrian age assignment are based upon tracing units across such a "bridge" near the St. Francis River about 50 miles north of the International Boundary (Cady, 1960, p. 542, 548-549). The series of axial synclines which are crossed on this field trip offers another possible bridge for a more detailed east-west correlation of the pre-Ottauquechee units.

Major structural units

The major structural units and features to be crossed on this trip are described on the following pages from west to east.

1) Hinesburg thrust -- The Hinesburg thrust marks the eastern limit of the Cambrian and Ordovician carbonate-quartzite assemblage, which is relatively unmetamorphosed and which includes distinctive and fossiliferous strata.

2) Georgia Mountain anticline and syncline -- In these south-plunging folds the Dunham Dolomite (Ꞓd) and Cheshire Quartzite (Ꞓc), which are dated units of the carbonate-quartzite assemblage, overlie a sequence consisting of the Fairfield Pond phyllite (Ꞓufp), White Brook dolomite (Ꞓuw), Pinnacle graywacke (Ꞓp), and Tibbit Hill volcanics (Ꞓpt). (Descriptive lithologic names are used here; formal usage is shown on the State Geologic Map.) Within this area the Tibbit Hill volcanics are the oldest rocks exposed, but they are probably interbedded with Pinnacle graywacke.

3) Enosburg Falls anticline -- The State Geologic Map shows a very complex pattern between the Pinnacle graywacke, the Tibbit Hill volcanics, and the Underhill phyllite. This complexity is due in part to folding, but much of it is due to lateral sedimentary intertonguing of these three units (see cross section A-A' of the State Geologic Map).

4) Cambridge-Richford syncline -- The Underhill schist, a silvery-green, white mica-chlorite phyllite or schist, occupies most of this area. The Underhill schist is bordered to the west by, and probably generally overlies, the Pinnacle graywacke, but it also intertongues with it. Similarly, the Underhill schist is bordered to the east by, probably generally overlies, and in part intertongues with albite schist, which is mostly shown as Hazens Notch Formation (Ꞓh) on the State Geologic Map. This syncline also contains several distinctive rock types including greenstone (Ꞓug and Ꞓup), limestone (Ꞓuw), and graphitic phyllite and slate (Ꞓuc, Ꞓs) whose detailed distributions are unknown.

5) Axial anticline of the Green Mountain arch -- The actual crest of the anticlinal arch is within rather coarse-grained porphyroblastic albite schist, both graphitic and non-graphitic, with

minor laminated quartzite interbeds. These grade eastward and upward into graphitic schist and quartzite with much less prominent albite.

6) Foot Brook syncline -- A shiny-green, paragonite- and chloritoid- bearing schist occurs in the core of the Foot Brook syncline. On the State Geologic Map this unit (Ɛufb) is shown as a facies tongue of the Underhill Formation, but it also has been tentatively correlated with the Stowe Formation (Ɛs) to the east (Albee, 1957 a, b).

7) Eastern limit of the Green Mountain anticlinorium -- East of the Foot Brook syncline is a generally homoclinal sequence of units - the Hazen's Notch (Ɛh), the Ottauquechee (Ɛo), the Stowe (OƐs), the Umbrella Hill (OƐu), the Moretown (Omm), the Shaw Mountain (Ss), the Northfield (DSn), and the Waits River (Dw). The pattern on the State Geologic Map indicates the existence of folds, most of them subparallel to the Green Mountain anticlinorium. The largest of these folds is the Worchester Mountain anticline within the Stowe Formation. The base of the Shaw Mountains marks a major unconformity which has been traced the entire length of Vermont.

Correlation across the axial anticline

The correlation of the Ottauquechee Formation on the east limb of the Green Mountain anticlinorium with Cambrian units west of the Hinesburg thrust is well established, but it is not clear whether the Ottauquechee Formation or units above it occur in the Foot Brook and Cambridge-Richford synclines. The pattern of units between the Hinesburg thrust and the Ottauquechee band on the east limb of the Green Mountain anticlinorium is explained on the State Geologic Map by a combination of folding and of sedimentary and metamorphic facies changes which involve only pre-Ottauquechee units. An alternative possibility (Albee, 1957 a,b) is that the shiny schist in the Foot Brook syncline (Ɛufb) correlates with the Stowe Formation, that the underlying greenstone (Ɛhg) with associated serpentinite correlates with the Belvidere Mountain amphibolite (Ɛhb), and that the intervening graphitic schist correlates with the Ottauquechee Formation, is continuous around the shiny schist, and extends northward into the area of the Ottauquechee Formation within a syncline south of Jay Peak (see State Geologic Map). It is also possible that the Ottauquechee Formation and higher units are present in the Cambridge-Richford syncline. The two bands of greenstone (Ɛug and Ɛup) may face each other across a syncline, with the Ottauquechee Formation lying above and between them, and correlate with the Belvidere Mountain amphibolite (Ɛhb). Alternatively, the graphitic phyllite (Ɛuc) and black limestone (Ɛuw) near North Cambridge may correlate with the Ottauquechee Formation. Such an interpretation is based in part on a generalized correlation of the albite schists of the axial region with

the Pinnacle graywacke. These suggestions do not deny extensive sedimentary facies changes such as are indicated by the State Geologic Map and would in fact require extensive facies changes. I simply wish to emphasize the need and importance of additional, very-detailed mapping and tracing of units within the axial synclines north and south of the Lamoille River.

Minor structural features

The axial anticline of the Green Mountain anticlinorium in northern Vermont is an arch of a well-developed schistosity. This schistosity is subparallel to bedding but is transverse to bedding in small folds with nearly east-west axes which are subnormal to the axis of the arch. The schistosity, bedding, and east-west folds are folded about nearly-horizontal, north-south fold axes which parallel the axis of the Green Mountain anticlinorium, and they are cut by a steep slip cleavage closely associated with the "Green Mountain" crinkles and folds. Within the axial region the relative time relations are consistent and well-displayed. To the west the schistosity steepens and the dominant foliation is a schistosity roughly axial planar to the major folds. This steep schistosity dominates in the phyllites within the Cambridge-Richford syncline and the Enosburg Falls anticline, and the bedding is sufficiently disrupted as to provide very little obvious guidance on the nature of the major folds in this area. The early east-west folds are rarely observable. Eastward from the axial region the schistosity also steepens; most outcrops contain steeply-plunging, fragmented folds, typically with a right-handed pattern. Nearly-horizontal, north-south folds are rarely observed in individual outcrops although they dominate the pattern of the major units. Transposition schistosity is common, and the bedding trend in individual outcrops within thin distinctive units is independent of the trend of the unit. Only in greenstone, amphibolite, and quartzite is a true bedding schistosity preserved over any distance. In the crest of the Worchester Mountain anticlinorium, amphibolite with nearly horizontal foliation is overlain by schist with nearly vertical schistosity.

In the easternmost part of the area to be covered by this trip, a steeply dipping, north-trending schistosity dominates; but two sets of folds subnormal to each other may be observed in some outcrops. The steep foliation is transected by a rather widely spaced cleavage which may be related to the doming in eastern Vermont. Konig and Dennis (1964, p. 43) infer from outcrops near Eligo Pond, which will be seen on this trip, that "...Green Mountain cleavage appeared to displace doming cleavage..." Albee (1968, p. 331) suggests that "...most of the deformational and metamorphic features in the rocks in northwestern Vermont along the Green Mountain anticlinorium are pre-Silurian, probably middle Ordovician." A detailed study of the minor structural features on either side of the unconformable base of the Shaw Mountain Formation would help to

resolve this problem but is made difficult by the generally poor outcrop in this horizon and the differing competency of the rocks.

Metamorphism

Most of northwestern Vermont lies outside the garnet zone within the biotite-chloritoid zone of metamorphism (see State Geologic Map; Albee, 1968). Higher grade rocks occur in elongate areas associated with the crests of the Green Mountain and Worchester Mountain anticlinoria and throughout much of northeastern Vermont. The route of this trip lies to the north of the garnet zone rocks in the Green Mountain anticlinorium, passes through the garnet and kyanite zone rocks in the Worchester Mountain anticlinorium, and extends into the garnet zone rocks of eastern Vermont. The higher grade rocks in the Worchester Mountains are extensively retrograded, and Albee (1968) has discussed evidence suggesting that the higher grade metamorphism is pre-Silurian and that the retrogradation occurred during Middle Devonian metamorphism responsible for the higher grade rocks in eastern Vermont.

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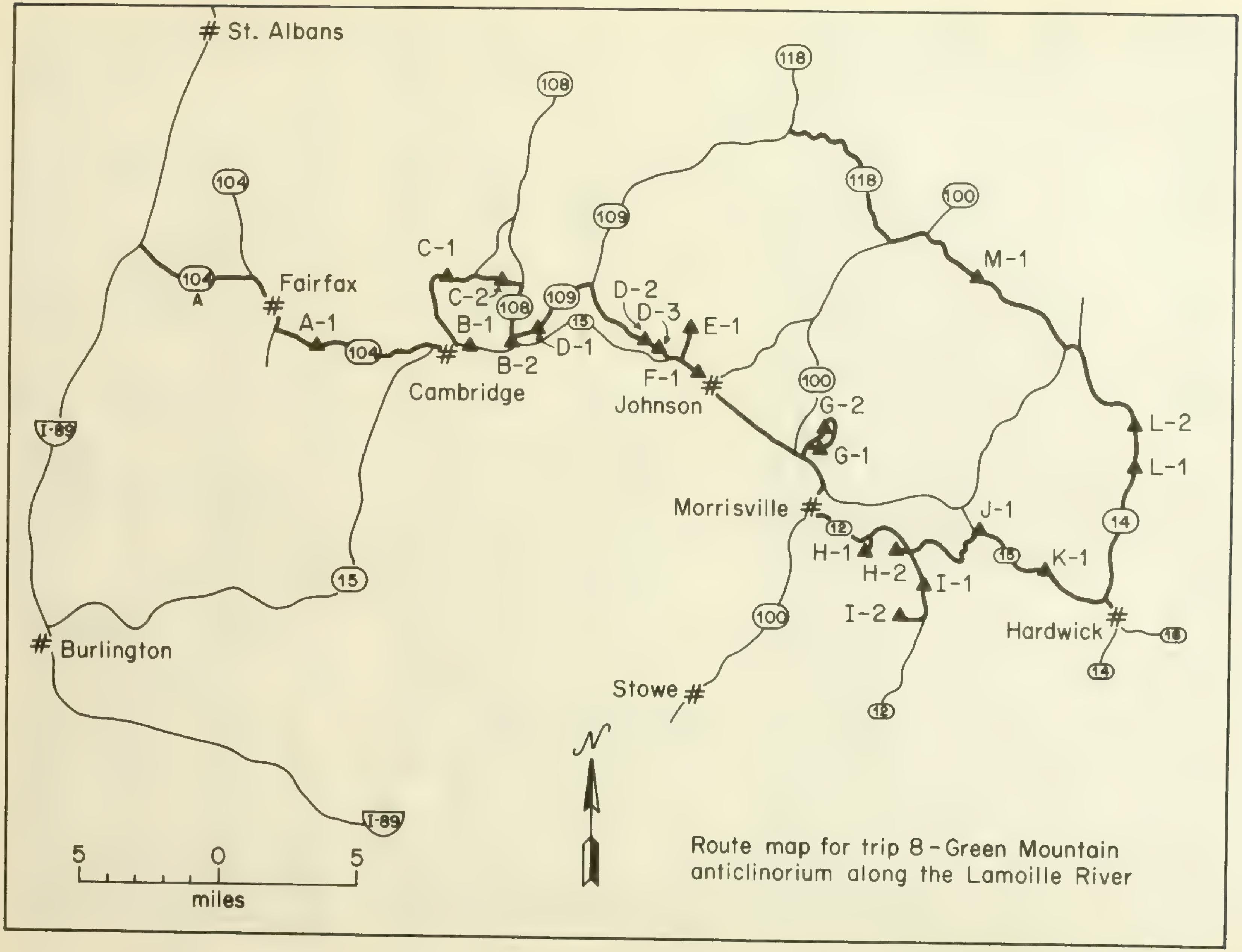

Route map for trip 8 - Green Mountain anticlinorium along the Lamoille River

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Road Log

Most of the stops described in this road log could be visited in one day by a small group, but a larger group would have to omit some stops. The choice of stops for the NEIGC trip will depend upon the weather and the size and interests of the group. The log is divided into segments or legs and the oddometer readings are reset at the start of each leg. The log has been written to be used in conjunction with the State Geologic Map, and no other maps are necessary. Only the general route and major stops are shown on the route map.

Start — Enter I-89 at exit 14, just east of the UVM campus and proceed north to the East Georgia exit. The route crosses Cambrian units of the Champlain Valley carbonate-quartzite sequence.

Leg A

0.0 — Leave I-89 at East Georgia exit, 17.6 miles north of Burlington, and proceed south on US-7.

0.3 — Y-intersection; proceed southeast (left) on Vt-104A.

1.3 — Georgia Mountain, directly ahead, consists of Cheshire Quartzite (Ɛc) and lies just east of the Hinesburg thrust.

1.9 — Large outcrop of Dunham Dolomite (Ɛd) on left.

2.0 — Caution - one-way railroad underpass. The overpass is located on the trace of the Hinesburg thrust.

2.2 — Outcrop of dark-colored Cheshire Quartzite on left. For the next several miles the route crosses a south-plunging anticline and syncline (see State Geologic Map) in which the Dunham Dolomite (Ɛd), the Cheshire Quartzite (Ɛc), the Fairfield Pond (Ɛu-Ɛp) and White Brook Dolomite members (Ɛuw) of the Underhill Formation, the Pinnacle Formation (Ɛp), and the Tibbet Hill volcanic member (Ɛpt) of the Pinnacle Formation occur in their "normal" stratigraphic sequence.

4.0 — Large roche moutonnée of Cheshire Quartzite (Ɛc). The bedding dips about 45° east, but the dominant foliation is nearly vertical.

4.8 — Y-junction with Vt-104; proceed south (right).

4.9 — Road cut on left in silvery-green phyllite, typical of Fairfield Pond Member (Ɛufp).

5.1 For next nine miles the trip crosses graywacke (Ꞓp) and volcanic (Ꞓpt) units of the Pinnacle Formation within the Enosburg Falls anticlinorium. The peculiar map patterns are inferred to be due both to folding and intertonguing relations.

5.5 Fairfax Village; continue on Vt-104 across the Lamoille River.

7.1 Y-junction; proceed east (left) on Vt-104.

8.1 Stop A-1 - Fairfax Falls
Caution - restricted visibility. Parking for 8-10 cars is available on left side of road.
Large road cuts in massive graywacke of the Pinnacle Formation (Ꞓp). Although biotite is present, clastic grains are evident and clasts up to 2 inches are present. The dominant foliation is a schistosity dipping steeply east; bedding is difficult to discern in such massive beds.

13.6 Junction - continue east on Vt-15. Sterling Mountain, directly ahead, is in the core of the Green Mountain anticlinorium.

14.1 Cambridge village.

14.5 "Wrong-way Bridge" over the Lamoille River. Local residents claim the engineers read the plans incorrectly with north and south reversed.

Leg B

0.0 Turn east (right) at north end of "Wrong-way Bridge".

0.5 Stop B-1 - Room for 8 cars on north side of road with clear visibility.
Typical silvery-green schist of the Underhill Formation (Ꞓu) is exposed in road cuts on the south side of the road. The dominant foliation is a near-vertical schistosity, transverse to quartz lenses and layers which form steeply south-plunging, right-handed folds. Flat-surfaced natural outcrops occur about 200 feet east on the north side of the road.

1.8 Bridge over Lamoille River. Continue straight.

2.0 Intersection in center of Jeffersonville. On the south side of the road, a war memorial has been carved in a large, complexly-folded outcrop of Underhill schist. The curved face and steps provide a spectacular exhibition in three-dimensions of complex minor folds with associated slip cleavage.

2.3 Junction of Vt-15 and Vt-108-109. Proceed north across bridge.

2.4 Stop B-2 - Parking on east side for 10 cars. The field northwest of the bridge contains excellent outcrops of silvery-green Underhill schist (Ꞓu), which exhibits considerable textural variation.

2.7 Junction; turn east (right) on Vt-109.

Leg C

0.0 Proceed north (left) from "Wrong-way Bridge" on blacktop road.

2.9 Road triangle with cluster of houses; turn east (right) on gravel road. Note that this intersection is about a mile west of the road shown on the State Geologic Map.

3.2 Stop C-1 - Park at top of hill. Outcrops of typical Tibbit Hill volcanic rocks (Ꞓpth) occur among the trees south of the road. These rocks are amphibolitic greenstones and contain coarse-amphibole, partially altered to actinolite and chlorite.

4.0 Turn north (left) on gravel road.

4.4 North Cambridge. Turn east (right) on gravel road.

4.8 Outcrop of Underhill schist on right.

5.2 Stop C-2 - Parking for about 6 cars on the right side along the turn; keep a flagman ahead. These outcrops include graphitic schist, quartzite, and limestone which are shown on the State Geologic Map as members (Ꞓuc and Ꞓuw) of the Underhill Formation within the core of the Cambridge-Richford syncline. On a lithologic basis it is conceivable that these rocks are correlatives of the Ottauquechee Formation (Ꞓo).

6.2 Turn south (right) on Vt-108.

8.2 Junction with Vt-109. Stop B-2, which is 0.3 miles south, has excellent outcrops of silvery-green Underhill schist (Ꞓu).

Leg D

0.0 Junction of Vt-108 and Vt-109. Proceed east on Vt-109.

0.9 Stop D-1 - Excellent road cut but no parking available for 0.2 miles. Quartz-muscovite-chlorite schist characterized by abundant albite porphyroblasts and containing graphitic interbeds.

The west-dipping schistosity has nearly-horizontal, north-trending crinkles and transects down-dip plunging tight folds. Such features will be seen in detail at the next stops.

3.6 The greenstone lense shown on the State Geologic Map (Ꞓhg) has been traced northward into the greenstone unit (Ꞓup) and into Quebec. Hence, the pattern shown on the State Geologic Map is known to be incorrect in detail, but the correct pattern is not known.

3.9 Junction - turn east (sharp right) off of Vt-109.

5.9 Large outcrop of massive, coarse-grained albite gneiss on the left. The eastward dip indicates its position just east of the crest of the Green Mountain axial anticline. Large open "Green Mountain" folds in large outcrops are visible from the road for the next several miles.

6.8 <u>Stop D-2</u> - Parking for 8-10 cars on right side. Drivers should look at this outcrop and then move on 0.5 mile to park on a rock point on the right side of the road. Park two cars abreast to allow room for 7 cars.

This series of outcrops includes graphitic and nongraphitic albite schist and gneiss as well as thin gray quartzites that display quite remarkable contortions. These rocks are typical of the Hazens Notch Formation near the crest of the anticlinorium where porphyroblastic albite is abundant. Note that some albite porphyroblasts are black due to included graphite, although little graphite remains in the matrix. Garnet crystals also occur within albite porphyroblasts, although no garnet occurs in the matrix of the schist.

In these outcrops a rock face subparallel to the road appears to show a bedding schistosity dipping about 45° east and cut by a near-vertical slip cleavage associated with a near-horizontal crinkle and open "Green Mountain" folds. However, rock faces sub-normal to the road show that the apparent bedding schistosity is parallel to the long limbs and axial planes of tight, down-dip plunging folds and actually transects bedding in the crests of these folds.

The time relationships can be discerned throughout these outcrops, but they are especially well shown at the second parking area on the cliff between the parking area and the river. The rock projection about 15 feet west of the oak tree is the crest of a "Green Mountain" fold, which plunges about 10° south. Crinkle axes parallel the larger fold axis. On both sides of the rock projection is a tight fold

in a gray quartzite layer. The fold axis is subnormal to that of the north-trending "Green Mountain" fold, passes entirely through the rock projection, and has been folded by the "Green Mountain" fold. This time relationship is consistently shown throughout the axial area of the Green Mountain anticlinorium.

7.7 Stop D-3 - Parking for 20 cars.
Similar relationships to that at the last stop are well displayed in a 1000 foot series of road cuts. The rocks are similar, but gray quartzites are more abundant.

8.3 Junction; turn east (left on Vt-15.)
The wooded ridge to the northeast has good outcrops of schist of the Foot Brook syncline (Ꞓufb). It is a highly-aluminous schist, commonly containing chloritoid, and has been variously correlated with both the Underhill Formation (Ꞓu) and the Stowe Formation (OꞒs).

8.8 Gravel road to north.

Leg E

0.0 Proceed north on gravel road.

1.4 Stop E-1
Silvery-green schist of the Foot Brook syncline (Ꞓufb) is exposed in the stream just north of the road.

2.8 Return to Vt-15.

Leg F

0.0 Proceed east on Vt-15.

0.5 Stop F-1 - Parking for 8 cars on right side of road beyond outcrop.
This outcrop is quite typical of those that might be seen in the Hazens Notch Formation (Ꞓh) for the next 6 miles eastward along the highway. They consist predominantly of graphitic schist and quartzite with minor nongraphitic quartzose schist and typically display pronounced sulfidic weathering. The rocks are fine-grained and do not typically contain biotite or porphyroblastic albite. The schistosity is typically steep, north-trending, and subparallel to the long limbs of folds in the quartzite beds. Abundant steeply plunging folds occur in the quartzite beds and in quartzose layers or pods in the schist. In this road cut it is possible to recognize "Green Mountain" folds, crinkles, and slip cleavage; but it is difficult to discern them in most natural outcrops.

1.1 Johnson Village.

4.4 The outcrops along the road contain an assemblage of rock types and structural features similar to those seen at the last stop.

5.8 Junction with Vt-100, continue east on Vt-15-100.

6.1 End of Leg F at blacktop road on north side of highway just east of small restaurant.

<u>Leg G</u>

0.0 Turn north (left) on blacktop road.

0.5 Y-junction, proceed east (right).

1.1 <u>Stop G-1</u> - Park along right side of road both east and west of Y-junction.
The large outcrop north of the road is in a narrow band of quartz-muscovite-chlorite-magnetite schist, (Ꞓhm) which occurs just below the Ottauquechee Formation (Ꞓo) at the position of the Pinney Hollow Formation (Ꞓph). Its silver-green color is in marked contrast to the graphitic-sulfidic rocks above it and below it. Although it extends in a straight north-trending band for about 10 miles, it typically contains very diverse bedding trends with steep folds and transposition schistosity. These structures appear to be internal to the unit. Similar bands to the west (see State Geologic Map) may be different units or may represent repetition of this unit by folding.

1.1 Proceed east (straight ahead).

1.4 Proceed north (left) on gravel road.

3.1 Proceed west (left) on blacktop road.

3.2 Continue straight ahead.

3.7 <u>Stop G-2</u> - Park on right. No hammers please.
The flat outcrop about 300 feet south of the road displays a wide variety of complex structural features reflecting the variation in competency of different lithologic types. The interbedded schist and quartzite in this outcrop is within the Ottauquechee Formation (Ꞓo) but is not typical of that unit.

4.1 Recross the Ꞓhm band at height of land.

5.4 Return to Vt-100-15.

<u>Leg H</u>

0.0 Proceed east on Vt-100-15.

1.6 East (right) on Vt-100 toward Morrisville.

2.6 Cross railroad tracks and turn east (left) into Morrisville.

2.7 Cross bridge and turn south (right).

2.9 Stoplight - turn east (left) on Vt-12.

5.1 Turn south (right) on Elmore Mountain Road.

5.4 Stop H-1 - Park on gravel road to east (left).
Abundant flat outcrops of amphibolite of the Stowe Formation (O€sg). These outcrops with their gently-dipping foliation are part of a broad band of amphibolite that dips under and reappears on the east side of the kyanite-grade schist on Elmore Mountain. At their contact flat-lying amphibolite is overlain by a coarse-grained muscovite-garnet-kyanite schist with a near-vertical, north-trending schistosity and very poorly preserved bedding. These garnet and kyanite zone rocks in the Worchester Mountain anticline have been extensively retrograded at a time postdating the formation of the dominant schistosity and folds.

Continue north on gravel road.

5.8 Turn east (right) on Vt-12. Numerous outcrops of the flat-lying amphibolite can be observed along the road on the north side of Elmore Mountain.

7.1 - 7.6 No room to park. Outcrops of coarse-grained schist of the Stowe Formation (O€s) containing garnet retrograded to chlorite.

7.7 Stop H-2 - Elmore State Park.
Outcrops of Stowe schist and amphibolite can be seen in the Upper Campground. Park at Campsite No. 5, drop about 200 feet south into the small stream, and go upstream about 600 feet to large outcrops of garnet-bearing schist. Small outcrops of amphibolite and a 15 foot glacial erratic of serpentinite may be seen downstream from the schist outcrop. A large outcrop of amphibolite is behind the rest room.

7.8 End of Leg H, Elmore Village.

Leg I 0.0 Side trip to kyanite-zone schist of the Stowe Formation. Proceed south on Vt-12 from Elmore Village.

1.1 Stop I-1 - Parking on side road to right.
Roadcut in compositionally-layered amphibolite of the Stowe Formation.

2.5 Outcrops of phyllite and "pinstripe" of Moretown member of Missisquoi Formation (Omm).

2.8 Stop I-2 - Go west (right) on dirt road; the condition of this road varies greatly, but it is always high-centered. May drive 0.5 mile west to edge of woods and park; then follow wood road an additional 0.4 mile across two streams to a point about 300 feet west of the second stream.
A prospect pit about 50 feet north of the wood road was dug for iron in a layer rich in ilmenite, kyanite, and chloritoid. In the area immediately north of the pit are good outcrops of kyanite-garnet-muscovite schist interbedded with coarse-grained garnet amphibolite. These rocks have undergone rather extensive retrogradation to aggregates of white mica, chloritoid, and chlorite. Return to Elmore Village.

Leg J

0.0 At Elmore Village turn east on gravel road.

3.8 Bridge over Lamoille River.

4.0 Stop J-1 - Turn east (right) on Vt-15 and park on right side. This outcrop of interbedded fine-grained quartzite and slate with biotite porphyroblasts is typical of the Moretown Member of the Missisquoi Formation (Omm). The bedding dips about 75° east; there are almost no minor folds, but there is a slightly more steeply-dipping cleavage.

Leg K

0.0 Proceed east on Vt-15.

0.9 Wolcott Village.

1.5 Bridge over Lamoille River. The abundant outcrops in the next mile are generally similar to those at the last stop.

3.2 Stop K-1 - Ample parking on left side with good visibility. The covered bridge is the only one still being used by a railroad and is characterized by its full-length ventilator. It was rebuilt and strengthened several years ago with donations collected in a state-wide drive.
The outcrop contains granulite, quartzite, and slate rather similar to that seen at the last stop; but a number of repetitions by larger folds are present. Both north-trending horizontal fold axes and steep fold axes subnormal to the horizontal axes are abundant; these fold relationships are similar to those seen in the axial region of the Green Mountain anticlinorium. Such relationships are rarely ob-

served in natural exposures. "Pinstripe" foliation has developed both parallel to and transverse to bedding. Garnet occurs locally in this outcrop although it is some distance west of the mapped garnet isograd.

6.0 Junction with Vt-14 - End of Leg K.

Leg L

0.0 Proceed north on Vt-14 along a valley typical of those associated with the Shaw Mountain and Northfield formations.

4.9 Stop L-1 - Parking on left; south end of Eligo Pond. Garnet-biotite phyllite and calcareous phyllite of Northfield Formation (D-Sn). There are several sets of crinkles, but the dominant folds plunge about 20°N 20°E and are overturned to the east.

6.2 Stop L-2 - Parking on left beyond outcrop. The roadcut contains green garnet phyllite of the Moretown Member (Omm) with gently north-plunging fold axes and crinkles with an associated slip cleavage. East of the road opposite the parking area is a large outcrop of biotite granulite, possibly a metavolcanic, which is part of the Shaw Mountain Formation (Ss). Several hundred feet further east in the trees is garnet-biotite phyllite of the Northfield Formation (D-Sn) similar to that seen at the last stop. The same structural features appear to be present in all three units, but a freeway cut would be very useful right here.

6.7 Large road cut in Northfield slate and quartzite with gently north-plunging folds overturned to the east. Garnet in this outcrop has been retrograded.

9.5 End of Leg L at blacktop road to west.

Leg M

0.0 Proceed west on blacktop road with directional sign pointing to North Wolcott.

0.4 Turn north (right) on gravel road at road triangle.

2.0 Continue west (left) at Y-junction with old school house.

5.1 Stop M-1 - Parking on left side of road at height of land. The most accessible outcrops are under the power line about 75 feet north of the road.
The Umbrella Hill conglomerate (Omu) crops out for about 25 miles as a thin band between the Stowe Formation (OЄs) and the Moretown Member (Omm). It contains subrounded quartz clasts and angular red, gray, yellow, and green

slate clasts up to 4 inches in a phyllitic matrix. The slate clasts are deformed into alignment with the schistosity of the matrix. Throughout its outcrop length the unit contains chloritoid plates in both clasts and matrix as well as quartz-kyanite veins. Both the chloritoid plates and the quartz-kyanite veins transect the schistosity. Quartz-kyanite veins are well exposed about 1000 feet south of the road, and several small veins occur in outcrops in the brush about 50 feet north of the power line.

6.4 Directly ahead can be seen the old asbestos workings near the top of Belvidere Mountain and the newer quarries lower down on the northeast side.

7.6 Outcrops on left are pebbly quartzite and schist of the Ottauquechee Formation (Ꞓo). From here north this unit contains abundant pebbly beds and differs considerably from its appearance to the south.

7.8 Junction with Vt-100; poor visibility, turn south (left).

7.9 Eden Mills Village. Road to north (right) leads to the asbestos mines.

9.5 Eden Corners. End of Leg M. Return to south via Vt-100 and Vt-15, or return to Burlington via Vt-118.

Leg N

0.0 Eden Corners - Proceed west on Vt-118.

4.7 - 5.7 New road cuts contain graphitic and non-graphitic albite gneiss similar to that seen in the axial region at the Lamoille River and contain similar structural features.

6.6 Junction Vt-109. Turn south (left) and follow Vt-109 for 15 miles to rejoin Vt-15 and return to Burlington.

Trip B-9

SUPERPOSED FOLDS AND STRUCTURAL CHRONOLOGY ALONG THE SOUTHEASTERN PART OF THE HINESBURG SYNCLINORIUM

by

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Terry Frank, and Thelma Barton
University of Vermont

INTRODUCTION

The regional geology of the Hinesburg synclinorium has been described by various authors; most notably Cady (1945, 1960, 1969), Welby (1961), Stone and Dennis (1964), and Stanley (1969, and this volume). The Centennial Geologic Map of Vermont (Doll et. al., 1961) is a representation of the state of knowledge of the synclinorium up to the time of its publication. More recent work carried out by various persons at the University of Vermont has greatly added to the knowledge of the structures and deformational history of the area.

It is the intention of this paper to bring together the attainments of the more recent work into an understandable and acceptable revision or alternate interpretation of the state geologic map.

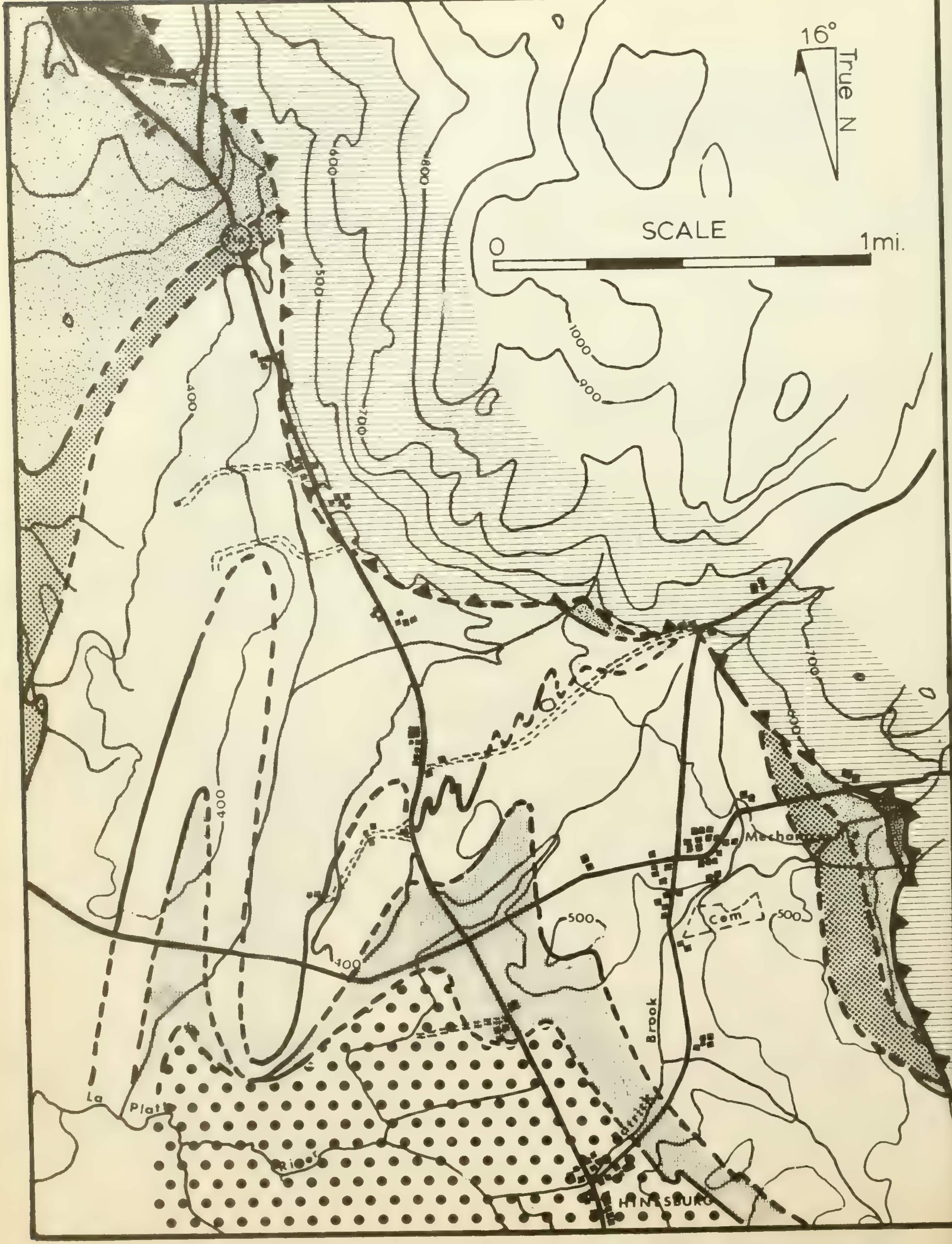
ACKNOWLEDGEMENTS

Recent work in the southern portion of the Hinesburg synclinorium has been carried out by several students at the University of Vermont. Information concerning the northwestern sector of the study area is largely drawn from unpublished reports of John Pratt, Thelma Barton and Barbara Gilman. Information from the western portion is taken from reports by Terry Frank and Thelma Barton. The eastern half of the area was studied by this author in conjunction with the preparation for a Master's Degree thesis at the University.

REGIONAL GEOLOGIC SETTING

For a brief account of the regional geologic setting the author refers you to Stanley and Sarkesian (Trip B-5) in this volume.

Figure 1


16°
True N
SCALE
0
1mi.
1000
900
800
700
600
500
400
Mechanicsville
Cem
Brook
La
Platt
River
HINESBURG

EXPLANATION

ORDOVICIAN	Lower	BROWNELL MTN. PHYLLITE MBR.
		BASCOM FORMATION
		CUTTING DOLOMITE
		SHELBURNE FORMATION
CAMBRIAN	Upper	CLARENDON SPRINGS DOLOMITE
		DANBY FORMATION
	Middle (?)	WINOOSKI DOLOMITE
	Lower	UNDIFFERENTIATED CHESHIRE QUARTZITE AND UNDERHILL PHYLLITE

Thrust fault, sawteeth on upper plate, Hinesburg fault

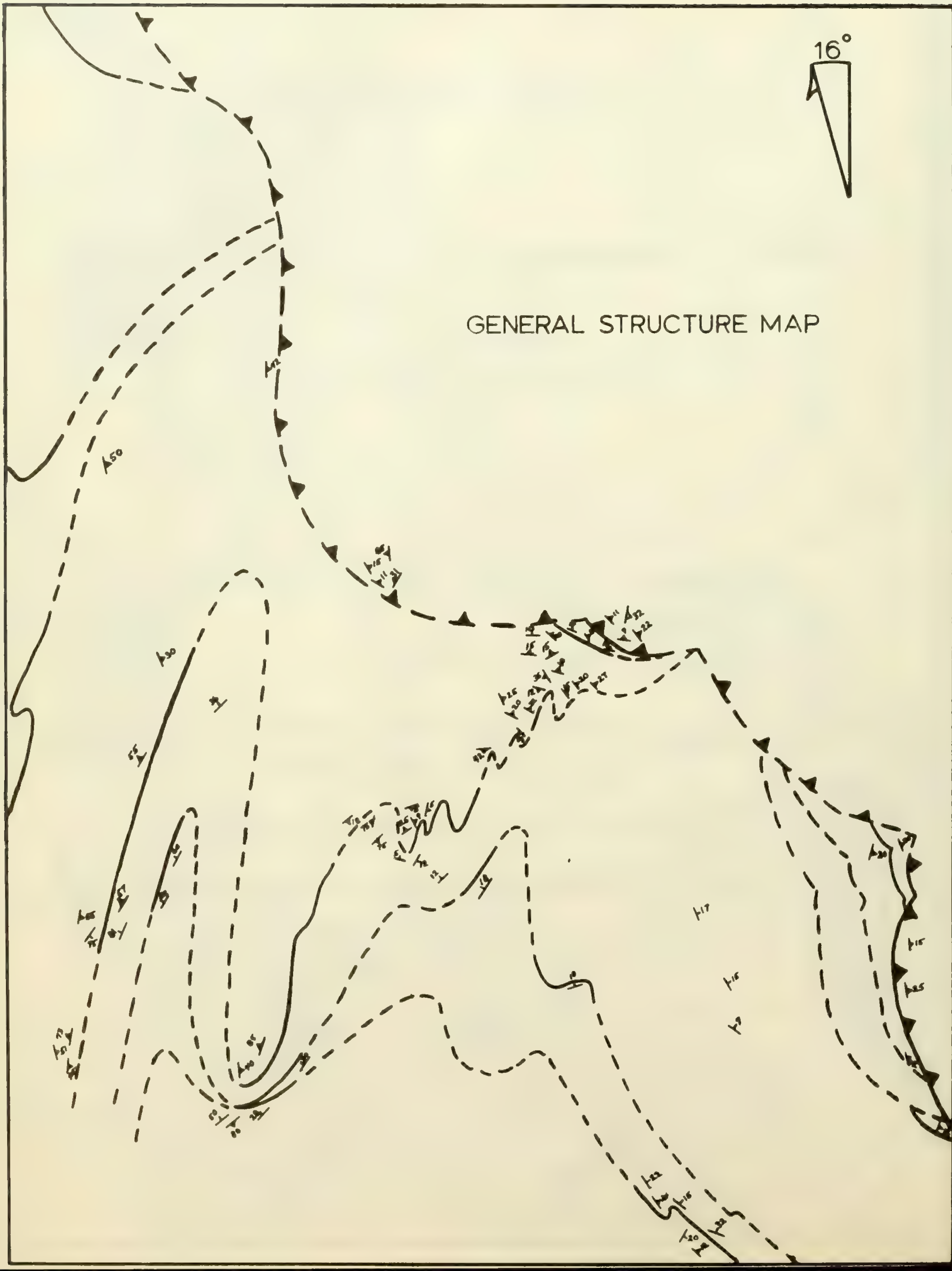
Formation contact, accurate

Formation contact, approximate

S-2 cleavage 20°   S-3 cleavage 10°

S-1 bedding 37

Figure 2



STRATIGRAPHY

The stratigraphy of the area is described in Table I and II in the paper by Stanley and Sarkesian in this volume (Trip B-5). Only the Winooski through the Bascom Formation in the Hinesburg area will be discussed here since recent mapping has concentrated on these units.

Lower Cambrian

Cheshire Quartzite - Typical Cheshire is a massive, very thick bedded white quartzite. The lower part is brownish weathering and is quite argillaceous and less massive. East of the Hinesburg thrust, the contact with the Underhill Formation is gradational and placed above the chloritic schists and phyllites and below the mottled gray argillaceous quartzites showing well developed slaty cleavage. The author has not mapped the contact in the Hinesburg area.

Dunham Dolomite - This formation is not present in the area of study but occurs extnsively to the south and west.

Monkton Quartzite - This formation also does not appear in the study area but outcrops extensively to the west in the upper plate of the Champlain thrust.

Middle Cambrian (?)

Winooski Dolomite - The Winooski consists of light gray to buff weathering dolostone, being gray to light pink or buff on the fresh surface. Thin phyllitic or siliceous laminae is sharp and tends to be marked by a distinct physical break. This contact zone, consisting of a thin phyllitic limestone with a closely spaced cleavage parallel to the contact is similar in appearance to a fault contact.

Upper Cambrian

Danby Formation - Beds of gray and brown crossbedded sandstone interlayered with beds of dolostone 1 to 2 feet thick are characteristic of the Danby Formation. The sandstones may be relatively pure massive white quartzites at some localities. A few thin layers of shale have also been observed near the top of the formation. In one locality (Stop #5), the basal Danby is an unusual boulder conglomerate made up of large blocks of sandstone,

dolostone, and quartzite in a sandy matrix. The attitude of bedding is more readily determined in the Danby than in the massive dolomites above and below. The contact with the overlying Clarendon Springs Dolomite appears to be gradational with the sandstones and interbedded dolostones of the Danby gradually giving way to the more dolomitic formation above.

Clarendon Springs Dolomite - The Clarendon Springs is a massive gray weathered dolostone, buff to gray on the fresh surface with a tendency to be coarsely crystalline. The most obvious feature of this unit is the presence of knots and segregations of quartz crystals standing out from the weathered surface. Some of the other dolostones in the section also show quartz knots but they are not as ubiquitous as in the Clarendon Springs. A few beds of calcareous sandstone stand out from weathered surfaces and are generally the only bedding indicator discernable in the monotonous dolostone section. Blue-black chert nodules are common near the top of the formation. The contact with the overlying Shelburne Formation is gradational and marked by a zone of mixture of the two with patches of Shelburne in depressions surrounded by more resistant dolostone.

Lower Ordovician

Shelburne Formation - The Shelburne is a massive dove gray weathered limestone and pink to white marble streaked with buff dolomitic stringers. There are also a few beds of sandy phyllite present. The Shelburne Formation is undoubtedly the most readily identifiable unit found in the Hinesburg area. The variety of rock types near the contact with the surrounding formations provides excellent structural markers of the several generations of folds in the area. The contact with the overlying Cutting Dolomite is usually sharp with sandy dolostone above and white marbles and gray limestones below.

Stone and Dennis (1964, p. 51) state that in the Milton area the Cutting "lies with destinct disconformity on the underlying Shelburne." In the Mechanicsville area, near the thrust contact, the Cutting seems to be absent entirely from the section placing the Bascom Formation on the Shelburne Formation. This could be due to a stratigraphic pinchout of the Cutting which begins in the area of St. George, Vermont. The other explanation would be that the Bascom-Shelburne contact is a thrust contact, the Bascom being dragged up along the Hinesburg thrust.

Cutting Dolomite - The Cutting is a massive whitish to light gray weathering dolostone, dark gray on the fresh surface with a tendency to be fine-grained. Large calcite crystals up to 1" across are common in some areas. The base of the formation is a thinly laminated cross-bedded, calcareous sandstone while the upper part contains black chert nodules. The contact with the overlying Bascom Formation was nowhere observed in the eastern half of the study area but Cady (1945, p. 543) states that there is no "apparent stratigraphic break."

Bascom Formation - This formation contains the widest variety of lithologies of any rocks in the Cambro-Ordovician section (Cady, 1945, p. 42). In the study area the Bascom is a blue-gray limestone with interbeds of buff to orange weathered dolomite and gray calcareous sandstone. Phyllitic laminae can be found in some of the limestone layers. The formation forms the lower plate of the Hinesburg thrust at the Mechanicsville exposure and appears discontinuously to the south.

Cady has more recently divided the Bascom Formation into the Brownell Mountain Phyllite Member and the typical Bascom (see Doll,et. al., 1961 and Cady, 1960, p. 539, footnote #7). According to Cady, the Brownell Mountain Phyllite is a calcareous phyllite in the upper part of the Bascom on the east limb of the Hinesburg synclinorium. During the course of recent field mapping a black calcareous phyllite has been found here and there along the Hinesburg thrust and in lens-shaped bodies on Brownell Mountain. Where the contact can be located within a few feet the change from limestone or dolostone to black phyllite is abrupt. Intermediate rock types between the limestone typical of the Bascom and the black phyllite have not been recognized. Two explanations are suggested for these relationships. First, the black phyllite may be a series of thrust slivers or older shales dragged up along the sole of the thrust plate and intermingled with slivers of allochthonous Bascom carbonates. This interpretation was suggested by Cady (1945, p. 567, 574, and Plate 10) in which the phyllite was correlated with the Skeels Corners Formation of Upper Cambrian age and formed the upper plate of the Muddy Brook thrust. Second, the black phyllite may be equivalent to the Hortonville and Walloomsac Formations of Middle Ordovician age that unconformably overlie older rocks in western New England. Subsequent movement along the Hinesburg thrust has plucked the black phyllites and mixed them with the other carbonate slivers which are found at such places as Mechanicsville near Hinesburg.

POSSIBLE UNCONFORMITIES

Evidence from field mapping seems to indicate that the Winooski Dolomite-Danby Formation contact is an unconformity. The evidence has been previously mentioned in the descriptions of the formations and is discussed under the description of Stop #5.

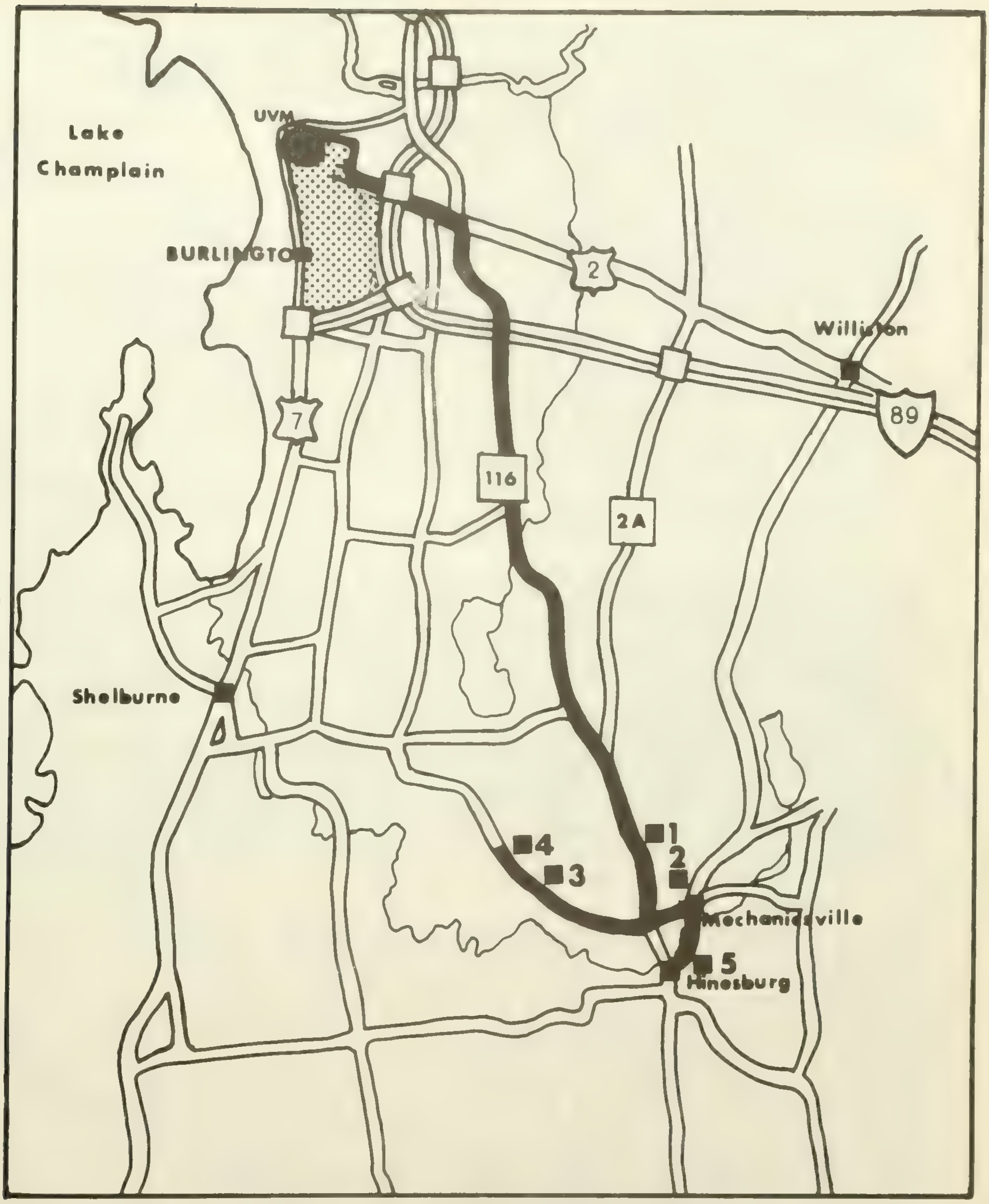
Another area which suggests an unconformity is the area east of the cemetary near Mechanicsville. Here the nondescript dolostones of the Cutting appear to rest on the dolostones of the Clarendon Springs with the Shelburne Formation notably absent, except for a small lens farther to the south. The contact between the two dolostones has not been directly identified but extensive exposure of the two formations makes it likely that the easily identified Shelburne occurs between them. The cause for the unconformity has not been determined. It may be a stratigraphic pinchout or a tectonic result of the nearby Hinesburg thrust; the Shelburne being tectonically squeezed out between the more resistant dolostones or absent due to imbricate thrusting.

The third area which suggests an unconformity is near the Mechanicsville exposure of the Hinesburg thrust. Sandy limestones of the Bascom Formation appear to rest unconformably on the Shelburne Formation with the Cutting absent. The actual contact of the two formations is covered by recent stream deposits. No boulders of dolostone could be found in the stream bed as might be expected if the Cutting were present. The presence of the Cutting is even somewhat doubtful in the area of the anticline depicted on Figure 1 near St. George to the northwest. Outcrops of Bascom and Shelburne occur quite close to each other there with no Cutting evident. Two explanations are possible for this unconformity. First, the Cutting is stratigraphically thinned and disappears to the northwest not reappearing from under the thrust at Mechanicsville. Second, the Bascom-Shelburne contact is a thrust contact at Mechanicsville, the Bascom being a thrust sliver dragged up along the sole of the thrust. However, if this were the case, one would expect a change in the bedding and first generation cleavage across the contact. This does not appear to be the case but does not entirely rule out the thrust hypothesis.

METAMORPHISM

A petrographic study of the Brownell Mountain Phyllite, the Fairfield Pond Member of the Underhill Formation as well as a thin phyllite in the Shelburne Formation

Figure 3



STOP MAP

SCALE

0 5 mi.

ROUTE

STOPS 1

was under taken to determine similarities and differences between the units. This was carried out with the intent of elucidating the modes of deformation, deformational histories, and grade of metamorphism relationships that may occur across the Hinesburg thrust fault.

Doll, et. al. (1961) describe the Fairfield Pond Member as a "greenish quartzite schist (quartz - sericite - albite - chlorite - biotite) and a sericite - quartz - chlorite phyllite." No biotite was observed in thin-sections from the Mechanicsville area in the Underhill Formation so it is not known at this time whether the Fairfield Pond Member reached the biotite isograd in this area. The phyllites of the lower plate were much more graphitic with the dominant phyllosilicate being very fine-grained muscovite. X-ray analysis is needed in order to more definitively locate the isograds. This will be undertaken in the near future.

FIELD TRIP STOPS

Stop 1 - Cheshire Quartzite on Rt. 116 north of Hinesburg. The first location is along Rt. 116 approximately 1 1/2 miles north of the village of Hinesburg. It is in the Cheshire Quartzite which forms the upper plate of the Hinesburg thrust in this area. At least two fold generations are present in this outcrop and possibly a third. The first generation slaty cleavage has been cut by a well developed spaced slip cleavage. The lineation formed by the intersection of the two cleavages may be systematically folded on a broader scale but lack of outcrop showing the lineation has not permitted detailed analysis. The Hinesburg thrust lies just to the west of the outcrop under recent deposits.

Stop 2 - Hinesburg thrust at Mechanicsville. This stop is the only exposure of the Hinesburg thrust in the Hinesburg synclinorium and is probably the finest in western Vermont. Here, the argillaceous Lower Cambrian Cheshire Quartzite rests upon the Lower Ordovician limestones and dolostones of the Bascom Formation. In at least two small areas the black phyllites, previously assigned to the Brownell Mountain Phyllite, can be seen below the thrust contact. The limestones of the lower plate and the argillaceous quartzites of the upper plate have been folded into minor northeast and southwest plunging anticlines and synclines. An exposure of the Brownell Mountain Phyllite just to the west of the thrust shows clear evidence for two generations of deformation with the development of a slip cleavage deforming the original slaty cleavage.

Figure 4

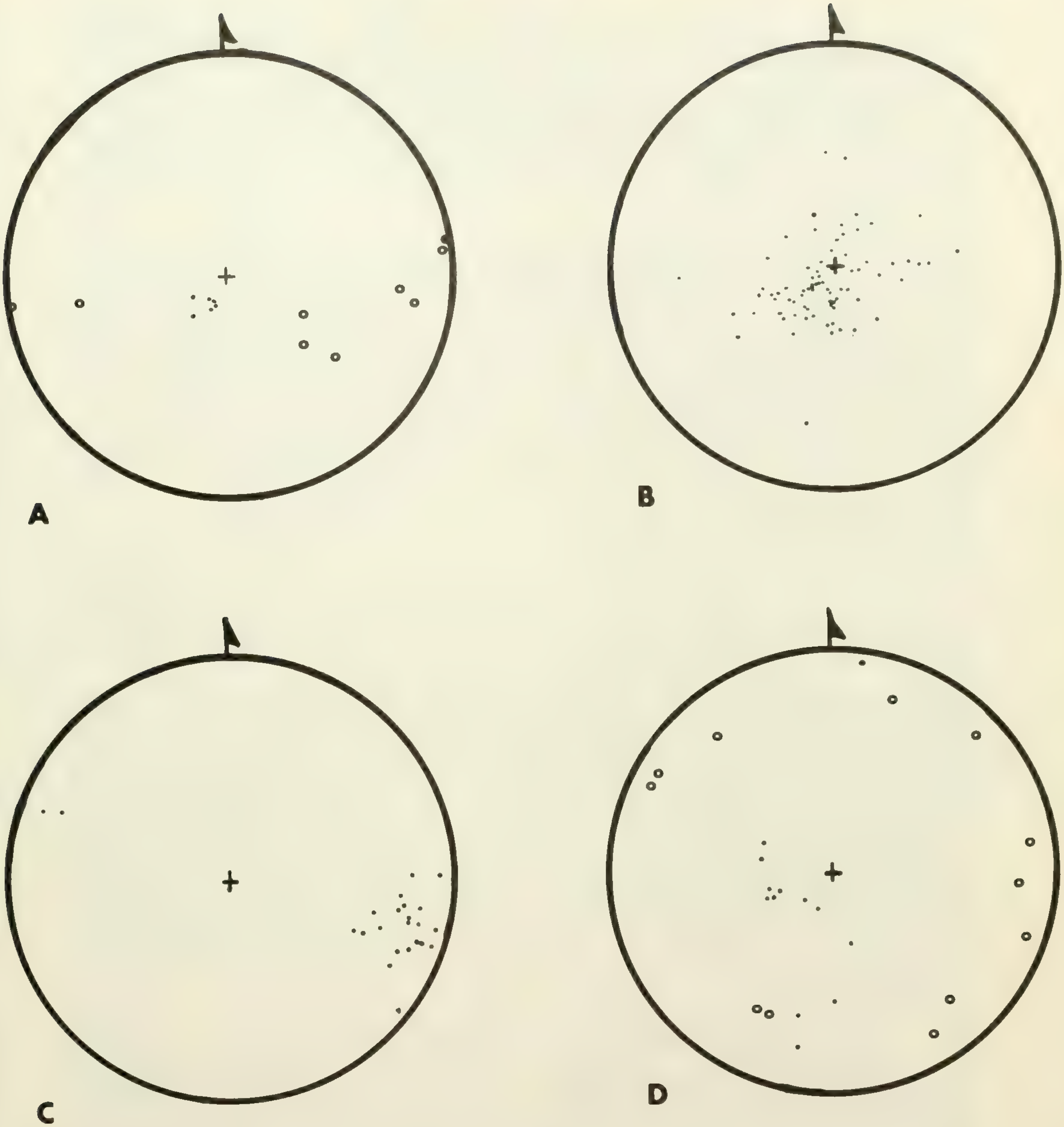


Figure 4 Lower hemisphere equal area projections of structures east of Rt. 116 in the town of Hinesburg, Vermont. A) poles to first generation closely spaced cleavage in Cheshire Quartzite (•) and second generation slip cleavage (∘) at Stop 1. B) 78 poles to first generation closely spaced cleavage in Shelburne, Bascom, and Brownell Mtn. Phyllite Formations (•). C) 22 lineations in Shelburne Formation formed by the intersection of bedding and first generation cleavage (•). D) poles to axial surfaces of folds in the Cheshire, Shelburne, Bascom, and Brownell Mtn. Phyllite Formations at Mechanicsville (•) and hinges of above folds (∘).

A short traverse down a nearby stream crosses lower Bascom sandy limestones and marbles of the Shelburne Formation. The intervening Cutting Dolomite, if present in this area, is very thin and not exposed due to recent cover (see above under Possible Unconformities). Dolomitic stringers in the white marbles give some indication of the bedding as well as display minor folds along the older cleavage.

Stop 3 - Ketcham's Pasture, East. As shown on the accompanying map and profile section, Figures 6 and 7, Ketcham's Pasture consists of an eastern portion, made up of an isoclinal anticline with an axial surface dipping steeply to the west, and a western series of more complexly deformed folds. Stop 3 is located in the transitional zone between the Shelburne and the Cutting Formations. The transitional zone consists of interbedded buff to tan, extensively jointed dolostones and blue-gray limestones with small sandy and shaly layers. Boudinage commonly accompanies the deformation of this unit and is displayed in an exposure approximately 1500 feet north of this stop.

At Stop 3, sandy layers in the limestone define bedding (S_1). Folding of bedding, as shown by equal area stereonet projection (Figure 8A), produces a hinge orientation of N9E at 7W. This F_1 folding produces an axial surface cleavage S_2. S_2 is well-developed and is recognized as a closely spaced (commonly less than one-sixteenth inch) cleavage transecting bedding. Figure 8B shows the orientation of the S_2 cleavage plane, N7E at 29E, and the S_2 axial surface, N16W at 32E. A second, younger fold generation is also recognized. F_2 folds deform S_2 and produce a slip cleavage S_3. S_3 is oriented at approximately the same strike as S_2 but at a steeper dip. As shown in Figure 8D, the S_3 cleavage plane is N10E at 80E and the S_3 axial surface is N13E at 64E. F_2 hinge measurements are scattered and are shown in Figure 8E.

At the southernmost outcrop at Stop 3, the two fold generations are well expressed. Hinges of both folds are present and refolding of S_2 cleavage is evident, producing a weakly developed S_3 cleavage. 150 feet to the north, a large antiformal F_2 hinge is associated with folding of the well-developed S_2 cleavage (Figure 5A). Poorly developed S_3 cleavage is oriented N60E at 48S at this outcrop. Further to the north, a series of F_1 folds are observed, again in the sandy layers in the limestone (Figure5B). F_1 hinge measurements from this outcrop are included in Figre 8C and define an axial surface oriented N5E at 15E and suggest a slip line of N80E at 15S.

Stop 4 - Ketcham's Pasture, West. At this stop in the western half of Ketcham's Pasture, F_2 deformation is less pronounced. Inside the gate, immediately to the west of the Ketcham residence, both fold generations can be seen. Figure 5C shows F_2 folds deforming S_2 cleavage at this location. 1000 feet to the northwest, however, at Stop 4, only the earlier fold generation is present.

Stop 4 is located near the contact between the Shelburne Formation and the transitional zone. Bedding in the Shelburne is folded around a hinge oriented N65E at 22S and the cleavage associated with this F_1 fold generation is not characteristic of the S_2 cleavage observed elsewhere in the area. It is a widely spaced slip cleavage resembling the S_3 cleavage at other localities and is oriented approximately N50E at 50S. Quartz filling of S_2 cleavage planes is common. In Figure 5D, sandy layers in the limestone show crinkle folds and display the widely spaced S_2 cleavage.

Slightly to the west of the main cliff at Stop 4 a basic dike, oriented N75W at 74S, intrudes the sedimentary sequence. It is typical of several such E-W striking, nearly vertical dikes found at Ketcham's Pasture and probably represents an event much younger than the most recent deformation at Ketcham's Pasture.

Stop 5 - Winooski Dolomite-Danby Formation Contact. This stop is unusual in that it has never been reported in the literature. We will cross Middle Cambrian (?) Winooski Dolostone and view the contact with the Upper Cambrian Danby Formation. In this area, as well as in all other areas where the contact has been seen, a sharp break separates the two formations. In this locality, however, the lower Danby contains a local boulder conglomerate; the large boulders being blocks of cross-bedded sandstone and quartzite and massive buff to brown dolostones set in a sandy matrix.

Whether the Winooski is Middle Cambrian in age or actually Lower Cambrian has never been determined directly. The complete absence of fossils has made it impossible to paleontologically date the formation in the Hinesburg area. In the St. Albans, Vermont, area, the Parker Formation underlies part of the Winooski there and has yielded Middle Cambrian fossils. Therefore, Stone and Dennis (1964) have assigned a Middle Cambrian age to the Winooski. Cady (1945) has placed the Danby in the Upper Cambrian while Stone and Dennis have correlated the Danby with the Woods Corners Group

Figure 5 Descriptions

A. Large F_2 antiform at Stop 3: The closely-spaced F_1 cleavage has been gently folded by F_2; the poorly developed, more widely-spaced F_2 cleavage is mostly easily seen in the lower left of the photo.

B. Tight F_1 folds at the northern end of Stop 3: A thin dolostone bed in the otherwise massive marble reflects the tight F_1 isoclinal folding.

C. F_1 and F_2 folds by the gate NW of the farm house: The light-colored dolostone bed is gently folded by F_1 which is then refolded by F_2 into tighter folds as seen in the lower left.

D. F_1 folds with quartz-filled cleavage planes at Stop 4: F_1 appears as crinkle folds in the more resistant dolostone beds with widely-spaced, quartz-filled cleavage planes. There is no evidence of F_2 in this area.

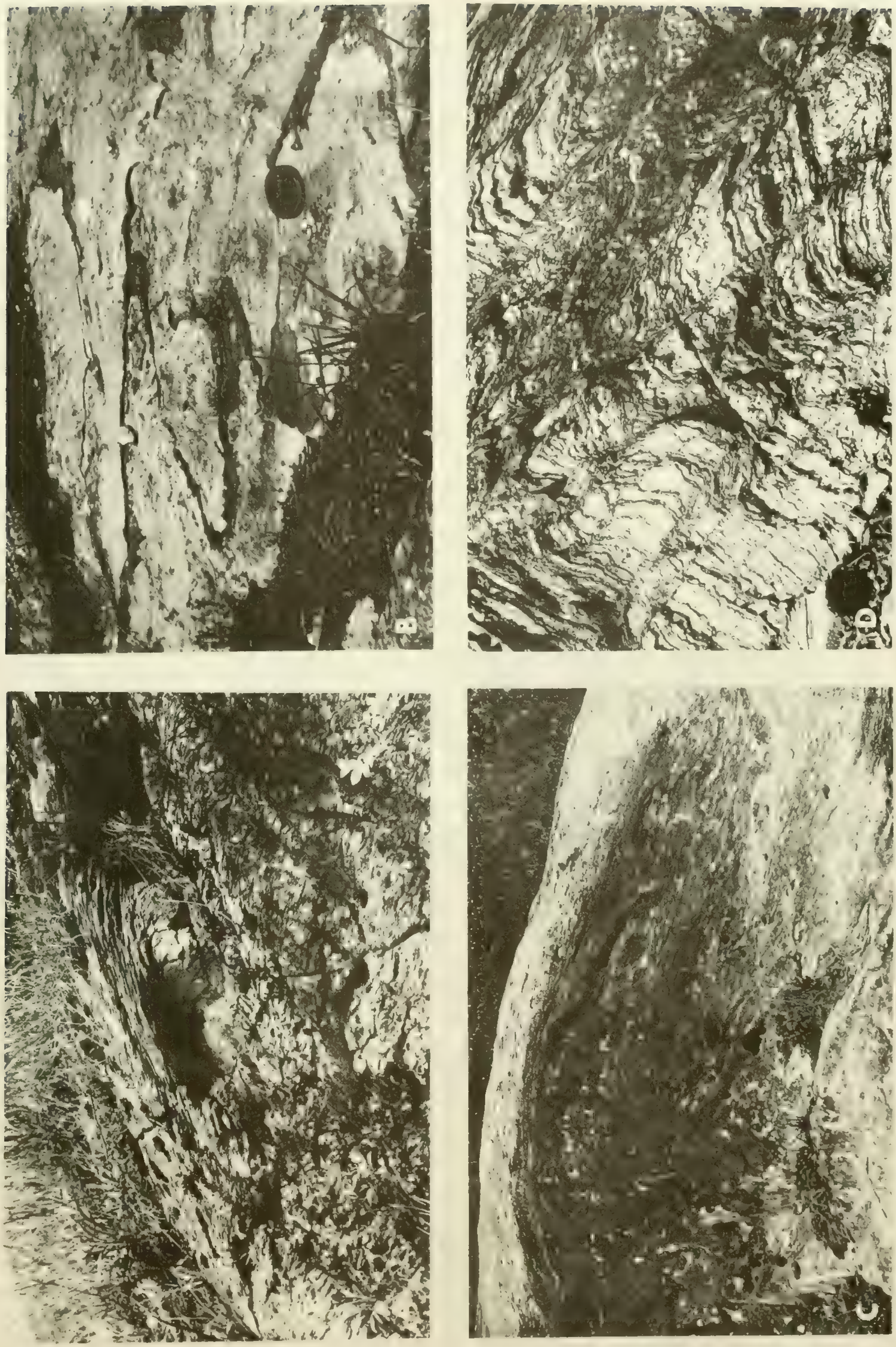
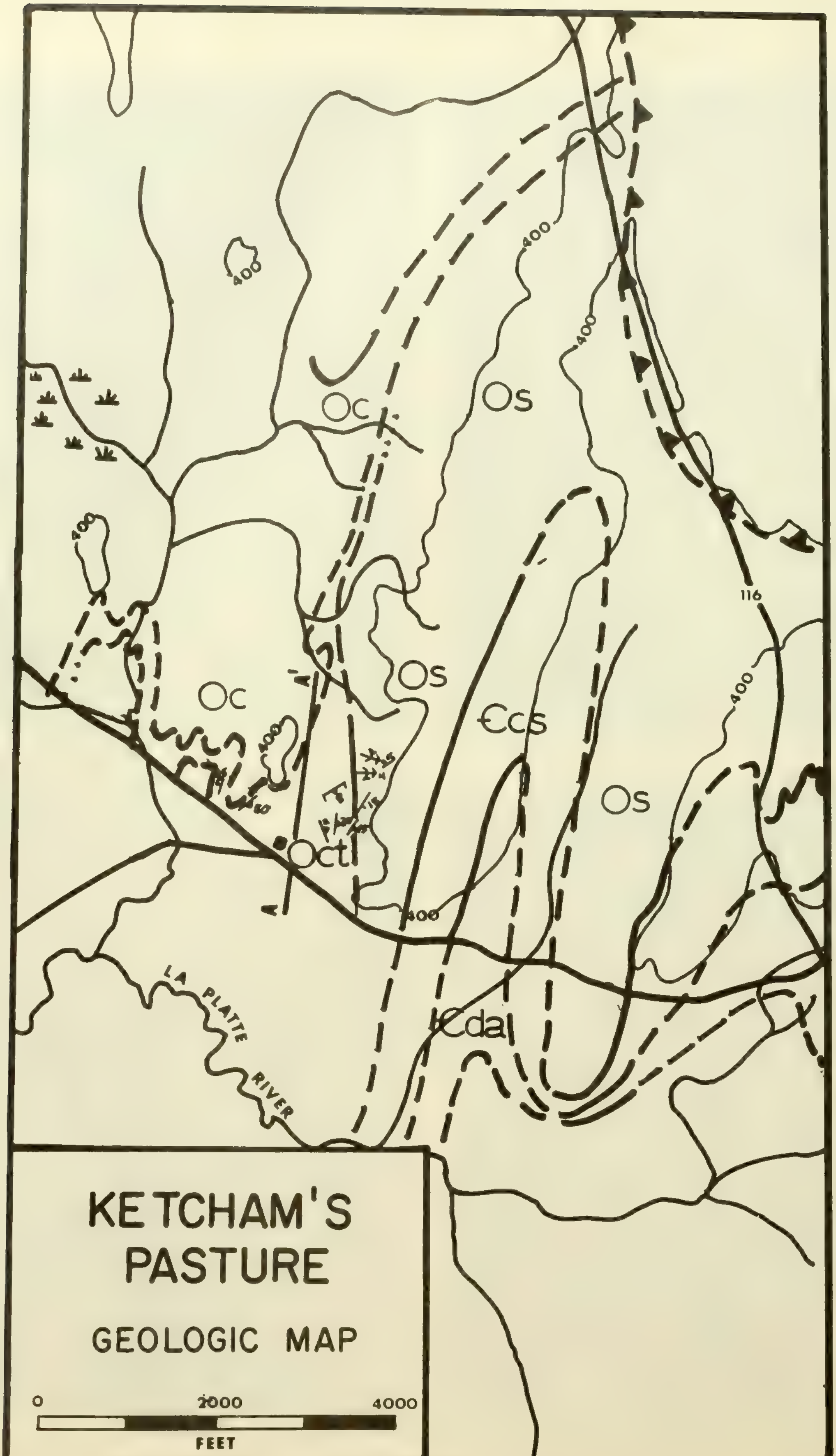


Figure 5


KETCHAM'S
PASTURE
GEOLOGIC MAP
0
2000
4000
FEET
Oc
Os
Ꞓcs
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Ꞓda
LA PLATTE RIVER
400
116
A
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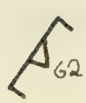
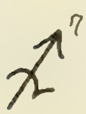
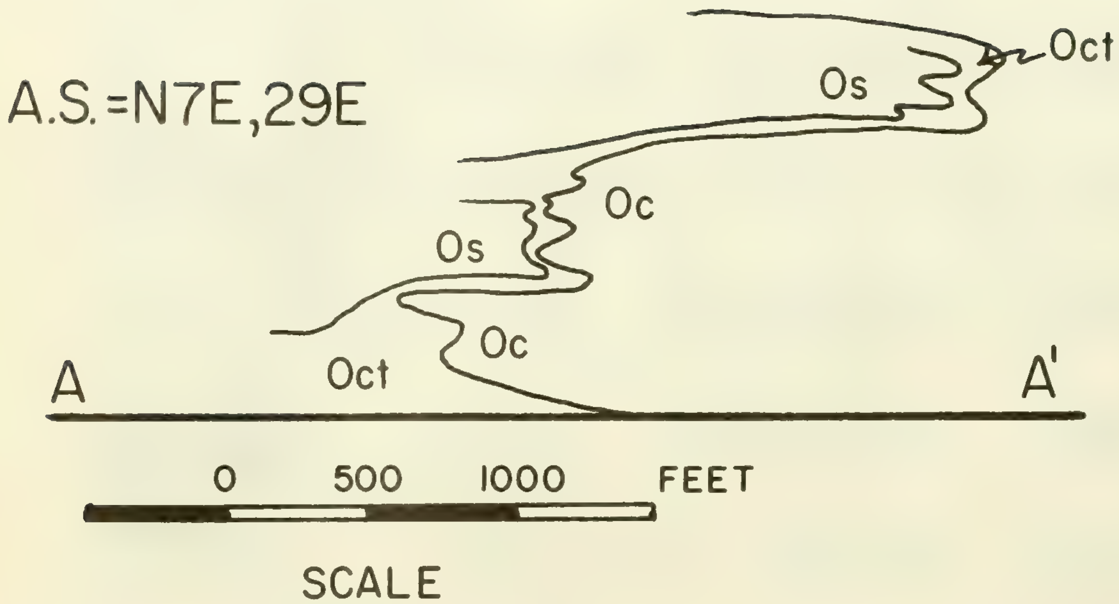
Figure 6

LEGEND

Oc	Cutting Dolomite
Oct	Cutting Dolomite - transitional
Os	Shelburne Formation
Ꞓcs	Clarendon Springs Dolomite
Ꞓda	Danby Formation
	Bedding
	F_1 Cleavage
	F_2 Cleavage
	F_1 Hinge - C.W. Rotation
	Formation Contact
	Questionable Contact
	Hinesburg Thrust Fault



Oc Cutting Dolomite

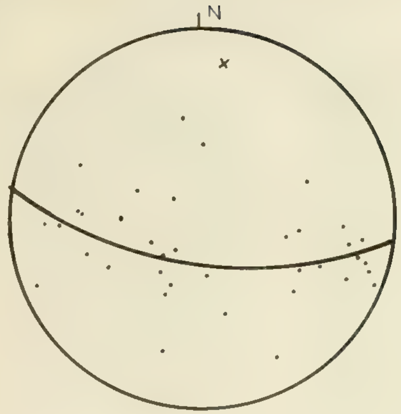
Oct Cutting Dolomite - transitional

Os Shelburne Formation

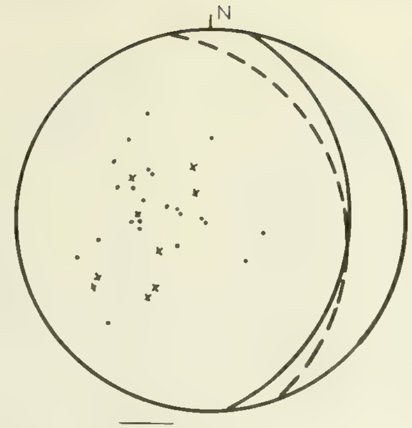
PROFILE SECTION

along the axial surface of F_1

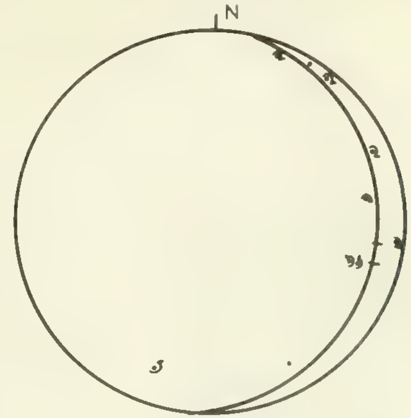
Figure 7



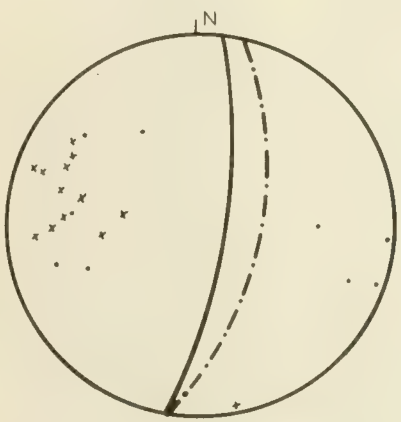
A. Poles to bedding: average hinge N9E,7N.



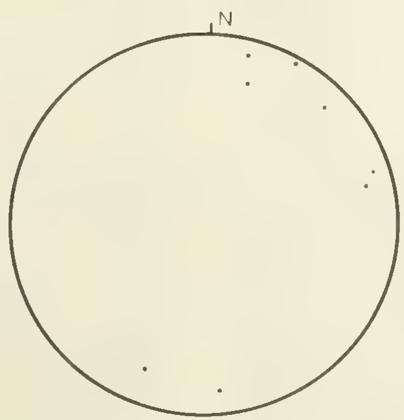
B. F_1 cleavage and axial surfaces: (·) poles to cleavage planes; (X) poles to axial surfaces; (—) cleavage plane, N7E,29E; (– –) axial surface, N16W,32E.



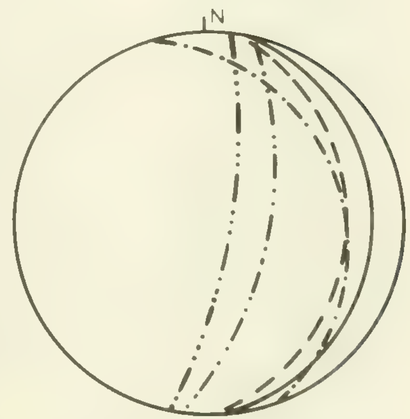
C. F_1 hinges: axial surface N5E,15E; separation arc 7°; slipline S80E,15E.



D. F_2 cleavage and axial surfaces: (·) poles to cleavage planes; (X) poles to axial surfaces; (—) cleavage plane N10E,80E; (–·–) axial surface N13E,64E.



E. F_2 hinges.



F. Summary net: (—) axial surface of F_1 hinges N5E,15E; (– –) F_1 cleavage N7E,29E; (–·–) F_1 axial surface N16W,32E; (–··–) F_2 axial surface N13E,64E; (–···–) F_2 cleavage N10E,80E.

Lower Hemisphere Equal Area Projections in the Area of Ketcham's Pasture.

Figure 8

of Middle Cambrian age. It appears possible that an unconformity may have existed, at least locally, between the Winooski and Danby Formations during Middle Cambrian time. This basal conglomerate seems to indicate a period of uplift and erosion preceding the deposition of the massive quartzites of the lower Danby. Positive identification of the boulders has not been possible as yet but hold the key as to the existence of the Middle Cambrian unconformity.

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Trip B10

LOWER PALEOZOIC ROCKS FLANKING THE GREEN MOUNTAIN ANTICLINORIUM

by

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Precambrian rocks designated the Mount Holly Complex (Doll _et al._, 1961) crop out over a large area in the southern and central Green Mountains of Vermont. The rocks of the Mount Holly Complex are probably correlative with the Grenville Series of the southeastern Adirondacks as described by Walton and deWaard (1963), and were metamorphosed and deeply eroded before the deposition of the Paleozoic (and perhaps also late Precambrian) rocks that now overlie them with profound unconformity.

All of the rocks have undergone severe Paleozoic deformation and regional metamorphism. The younger rocks on the west limb of the Green Mountain anticlinorium are in the biotite zone and those to the east are mainly in the garnet zone. The fabric and mineralogy of the rocks of the Mount Holly Complex have been strongly affected by the Paleozoic deformation and recrystallization. These effects are generally most pronounced at or near the unconformity separating the Mount Holly from the younger units, where the original textures and mineral assemblages of the Mount Holly are locally almost obliterated by later recrystallization and development of a penetrative schistosity. The localization of these features is probably related in part to weathering on the Precambrian erosion surface and in part to fluids derived from the prograde metamorphism of the overlying sediments.

The younger rocks on the west flank of the Green Mountain anticlinorium constitute the Champlain Valley Sequence as outlined in Table 1. This sequence includes about 1500 feet of basal clastics of earliest Cambrian (and perhaps also late Precambrian) age, overlain by about 3500 feet of Cambrian and Lower Ordovician carbonate rocks with minor intercalations of phyllite and quartzite. On the east flank of the anticlinorium, however, the Mount Holly is overlain by a much thicker sequence of schists

and phyllites containing several metavolcanic units. This sequence, the Eastern Vermont Sequence as outlined in Table 2, also includes possible late-Precambrian, Cambrian and Ordovician rocks. The allochthonous Taconic Sequence now located west of the Green Mountain region is outlined in Table 3. The Taconic Sequence is similar in many ways to the Eastern Vermont Sequence but contains more carbonates and less evidence of volcanic activity. It was presumably deposited in or near what is now the Green Mountain region, but the original site of deposition is now foreshortened by subsequent deformation. The narrow septum of younger rocks extending north-northwest from Pico Peak (Figure 1) is of considerable interest in this regard. A tentative correlation between the rocks of the three main sequences of southern and central Vermont is given in Figure 2.

The primary purpose of this field trip is to acquaint interested geologists with the stratigraphic sequence on the east limb of the Green Mountain anticlinorium in south-central Vermont. Most of the outcrops visited are on U.S. Highway 4 between Sherburne Center and Bridgewater Corners along the valley of the Ottauquechee River. One stop (Stop 1) will be made near Rutland, however, to see the upper part of the Cheshire quartzite and the basal units of the Rutland (Dunham) dolomite. The reason for this stop is to compare this sequence with a strikingly similar one in the upper part (Plymouth Member) of the Hoosac Formation near Plymouth, Vermont (Stop 14).

The unconformity between the Tyson Formation and the underlying Precambrian basement will be visited at Sherburne Center followed by representative exposures of the overlying Tyson, Hoosac, Pinney Hollow, Ottauquechee, Stowe and Missisquoi Formations. The route of the excursion will pass through the type exposures of the Ottauquechee Phyllite and Pinney Hollow Schist as originally defined by E.L. Perry (1929).

The geologic sketch map (Fig. 1) is modified from the Centennial Geologic Map of Vermont (Doll et al., 1961) on the basis of recent geologic investigations by the author and by P.H. Osberg (1959, and later communications). All outcrops to be visited are in the Rutland and Woodstock quadrangles (U.S.G.S. 15' series) for which there are published maps by W.F. Brace (1953) and by Chang, Ern, and Thompson (1965), respectively. Other publications pertinent to the area of the excursion are those of Thompson (1959, 1967), Zen (1961, 1964) and Osberg (1959) for the area near Rutland on the west flank of the Green Mountains, and those of Osberg (1952) and Ern (1963) for the central Green Mountain area and the region immediately to the east.

The route of the excursion is also covered by new topographic maps of the U.S.G.S. 7 1/2' series. The route of the excursion passes through parts of the Rutland, Chittenden, Pico Peak, Killington Peak and Plymouth quadrangles in that order.

The arguments for the dating of the Eastern Vermont Sequence have been summarized by Chang et al. (1965). The dating of the Champlain Valley Sequence has been reviewed by Theokritoff and Thompson (1969) who also summarize recent findings on the dating of the Taconic Sequence. The correlations implied by Figure 2 are at least consistent with the paleontologic data now available. One of the principal differences between Figure 2 and the correlations of other authors (Zen, 1967, Plate 2, for example) is in the dating of the Tyson, Hoosac and Pinney Hollow Formations relative to the lower part of the Champlain Valley Sequence. The evidence for the revisions proposed here is admittedly circumstantial and is based in part on the intriguing similarity between the sequences to be seen at Stop 1 and at Stop 14 on this excursion. A second major factor influencing the construction of Figure 2 is the presence of iron ores at the contact between the dolomite member of the Tyson Formation and the overlying albite schists of the Hoosac Formation as seen at Stop 7. This is taken as evidence for a period of subaerial erosion and correlated with similar occurrences reported by Booth (1950) and others at the contact between the White Brook Dolomite and West Sutton Formation in northwestern Vermont and southern Quebec.

Table 1

Cambrian and Lower Ordovician rocks of the Champlain Valley Sequence near Rutland, Vermont (Modified from Thompson, 1967)

<u>Bascom Formation</u> (Ob) Lower Ordovician

Interbedded calcite marble and dolostone. (350-400')

<u>Shelburne Marble</u> (Os) Lower Ordovician

White calcite marble. (250')

<u>Clarendon Springs Formation</u> (€cs) Upper Cambrian

Upper member: Cherty dolomite. (150-200')
Sutherland Falls member: White calcite marble, dolomitic curdling. (50-100')
Lower member: Gray calcitic dolomite, cross-bedded sandy dolomite. (200-250')

<u>Danby Formation</u> (€d) Upper Cambrian

Interbedded vitreous quartzite and cross-bedded sandy dolomite. (50-150')

<u>Winooski Dolomite</u> (€w) Middle Cambrian

Varicolored dolomites, minor dolomitic quartzite and schistose quartzite. (300-400')

<u>Monkton Quartzite</u> (€m) Lower Cambrian

Quartzite, schistose quartzite and feldspathic quartzite interbedded with varicolored dolomites and minor phyllite. (300')

<u>Rutland (Dunham) Dolomite</u> (€r) Lower Cambrian

Gray and yellow weathering dolomites, thin siliceous partings. (900')

Table 1 Continued

<u>Cheshire Quartzite</u> (€c) Lower Cambrian

Mainly vitreous quartzite, gray to black quartzose phyllite in lower part. (1000-1600')

<u>Dalton Formation</u> (€dt) Probably Lower Cambrian

Schistose graywacke, conglomerate, minor phyllite; discontinuous dolomite or sandy dolomite near top. (50-300')

Table 2

Cambrian and Ordovician rocks of the Taconic Sequence near Rutland, Vermont (Modified) from Zen, 1961; Thompson, 1967; Theokritoff and Thompson, 1969)

<u>Pawlet Formation</u> Middle Ordovician

Graywacke and interbedded black slate. (700')

<u>Indian River Slate</u> Middle Ordovician

Red and blue-green slate. (200')

<u>Poultney Slate</u> Lower Ordovician

Mainly thin-laminar, siliceous slates, minor limestones near base. (600')

<u>Hatch Hill and West Castleton Formations</u> Lower to Upper Cambrian

Black slate, dolomitic quartzite, minor limestone. (500')

<u>Bull Formation</u> Lower Cambrian

Mettawee Slate: Purple and green slate, thin limestone conglomerate near top; green phyllites or schists (St. Catharine Formation) in eastern Taconics.
Bomoseen Graywacke: Graywacke, minor slate and quartzite; albitic phyllites with quartzite, dolomite and limestone in upper part (Netop Formation of Thompson, 1967) in eastern Taconics. (600')

Table 3

Cambrian and Ordovician rocks of the Eastern Vermont Sequence (Modified from Chang *et al.*, 1965)

Missisquoi Formation (Om) Lower and Middle Ordovician

Cram Hill Member: Black, sulfidic schist, schistose quartzite. (250')
Barnard Volcanic Member: Biotite gneiss, hornblende gneiss, amphibolite (2500')
Moretown Member: Quartzite and quartz feldspar granulite with thin micaceous partings producing a "pinstripe" texture. (2000')
Whetstone Hill Member: Gray to black phyllite, micaceous quartzite, amphibolites, coticule and quartz-garnet-magnetite rock. Minor pinstripe quartzite. (2000')

Stowe Formation (O-Ꞓs) Ordovician or Cambrian

Quartz-sericite-chlorite schist with garnets and biotite abundant locally. (1500')

Ottauquechee Formation (Ꞓo) Lower to Upper Cambrian

Black, sulfidic phyllite or schist, quartz-sericite-chlorite-schist with garnet and biotite. Vitreous quartzites, some of which are carbonaceous, occur as beds to ten feet thick near base, but thinner and less abundant above. Greenstones and actinolitic greenstones occur locally. (3000')

Pinney Hollow Formation (Ꞓph) Lower Cambrian

Quartz-sericite-chlorite schist-biotite and garnet abundant locally and green chloritoid phyllite abundant in lower part. Some layers of chloritoid phyllite have a faint purplish color due to hematite. Greenstone and actinolitic greenstone in upper part. (2000')

Table 3 Continued

<u>Hoosac Formation</u> (€h) Lower Cambrian

Albitic schists, schistose feldspathic quartzites, carbonaceous schists near top. Middle and upper part of formation (Plymouth Member) contains vitreous quartzite, dolomite, dolomite breccia and dolomite with carbonaceous partings. (1250')

<u>Tyson Formation</u> (€t) Probably Lower Cambrian

Upper member: Dolomite characterized by lenses of magnetite or hematite at or near top. These were formerly mined as iron ore and may be a metamorphosed <u>terra rosa</u>. (200')
Middle member: Quartzite and pebbly quartzite interbedded with calcitic and dolomitic marbles and black phyllite. (250')
Lower member: Conglomerate and schistose graywacke. (350')

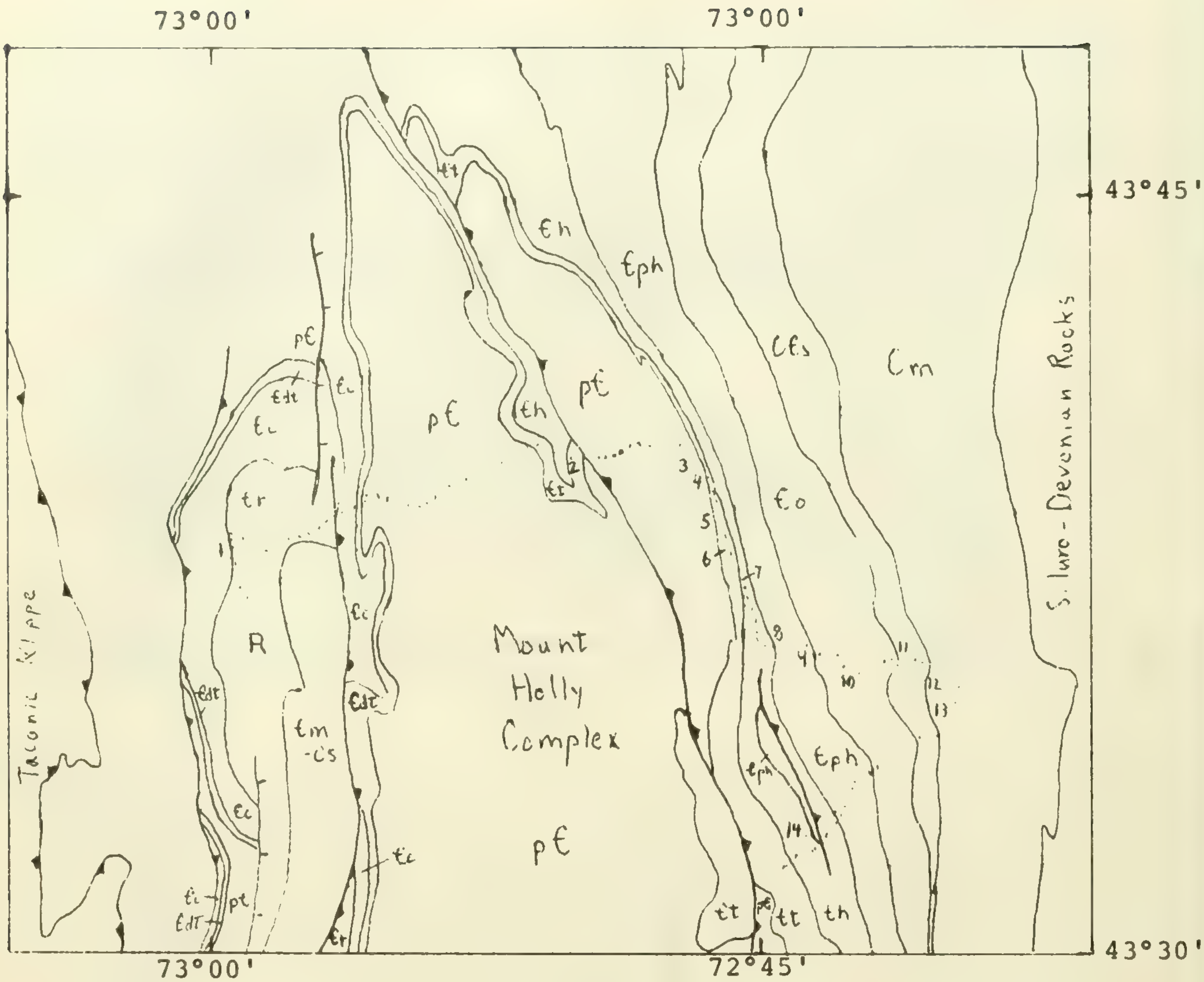

Figure 1. Geologic sketch map of a part of south-central Vermont, modified from Doll *et al*. (1961). Symbols as in Tables 1 and 3. Dotted line is route of excursion starting at Rutland (R) with stops numbered as in road log. Scale 1:250,000.

<table>
<tr><th></th><th>Champlain Valley</th><th colspan="2">Taconic Range</th><th>Eastern Vermont</th></tr>
<tr><td rowspan="3">Middle Ordovician</td><td rowspan="2">Ira Fm.</td><td colspan="2">Pawlet Fm.</td><td rowspan="4">Missisquoi Fm</td></tr>
<tr><td colspan="2" rowspan="2">Indian River Slate</td></tr>
<tr><td>Baker Brook Volcanics</td></tr>
<tr><td rowspan="2">Lower Ordovician</td><td>Bascom Fm.</td><td colspan="2" rowspan="2">Poultney Slate</td></tr>
<tr><td>Shelburne Marble</td><td rowspan="2">Stowe Fm.</td></tr>
<tr><td rowspan="3">Upper Cambrian</td><td rowspan="2">Clarendon Springs Fm.</td><td colspan="2" rowspan="4">Hatch Hill Fm.</td></tr>
<tr><td rowspan="6">Ottauquechee Fm.</td></tr>
<tr><td>Danby Fm.</td></tr>
<tr><td rowspan="3">Middle Cambrian</td><td rowspan="2">Winooski Dol.</td></tr>
<tr><td colspan="2" rowspan="3">West Castleton Fm.</td></tr>
<tr><td rowspan="2">Monkton Qte.</td></tr>
<tr><td rowspan="5">Lower Cambrian</td></tr>
<tr><td rowspan="2">Rutland Dol.</td><td colspan="2">Mettawee Slate</td><td>Pinney Hollow Fm.</td></tr>
<tr><td rowspan="2">Bomoseen Gray-wacke</td><td rowspan="2">Netop Fm.</td><td rowspan="2">Hoosac Fm.</td></tr>
<tr><td>Cheshire Qte.</td></tr>
<tr><td>Dalton Fm.</td><td colspan="2">Conglomerates on Bird Mtn.?</td><td>Tyson Fm.</td></tr>
</table>

Figure 2. Stratigraphic correlation of some Cambrian and Ordovician rocks in south-central Vermont.

Road Log for Trip B10

Starting point is the municipal parking lot opposite the Hotel Bardwell (near City Hall), Rutland, Vermont (Rutland 7 1/2 minute quadrangle).

Mileage

0.0	Parking lot, proceed north via Merchants Row and Grove Street.
1.4	Enter Chittenden 7 1/2 minute quadrangle and pass golf course on left with exposures of Rutland (Dunham) Dolomite.
1.7	Cross East Creek (Lower Cambrian fossils upstream, Theokritoff and Thompson, 1969, Stop 2).
2.2	Stop 1: Power line crossing. Walk west over exposures of dolomite breccia at base of Rutland Dolomite. Topmost beds of Cheshire Quartzite are exposed on hill 0.15 miles W.
2.6	Turn E. (right) on McKinley Lane and follow to U.S. Highway 7.
3.7	Turn N. on U.S. 7.
3.9	Turn E. on Post Road.
4.9	Rutland Dolomite on left.
5.6	Turn right (E.) on Park Lane and follow to U.S. Highway 4.
6.5	Turn left (N.) on U.S. 4. Outcrops just N. on right are part of Mount Holly Complex.
6.9	Mendon Village. A fairly complete section from the Mount Holly Complex, through the Dalton Formation, into the lower part of the Cheshire Quartzite is exposed about one mile north of here at the W. base of Blue Ridge Mountain. This is the type locality for the Mendon Series of C.L. Whittle (1894, p. 408-414). Proceed E. on U.S. 4 past outcrops (7.3-8.0) on right of gneisses, schists, quartzites and calc-silicate marbles of Mount Holly Complex.
9.8	Enter Pico Peak 7 1/2 minute quadrangle.

11.2 Beaver Pond on left. In notch one mile northwest (accessible via Elbow Rd.) are extensive exposures of dolomites now assigned to the Tyson Formation, and also, on the hill E. of the notch, of the lower part of the Hoosac Formation.

11.7 Pico Ski Area on right.

12.3 Stop 2: Summit of Sherburne Pass. Outcrop S. of road and ledges on Deer Leap Mountain to N. are gneissic conglomerate and metagraywacke now assigned to the Tyson Formation. These rocks are in the Pico syncline, a narrow, east-dipping septum containing rocks of the Eastern Vermont Sequence. The eastern boundary of the septum is probably a major thrust fault. Outcrops E. of pass are in Mount Holly Complex.

13.8 Junction with Route 100, proceed E. on U.S. 4 past access road to Killington Ski Area. Outcrops of gneisses of Mount Holly Complex appear sporadically on right side of road over next mile and a half.

15.5 Start of long road cut on right in Mount Holly Complex.

15.7 Stop 3: Unconformity at base of Tyson Formation is exposed, though perhaps not convincingly, near east end of cut about 0.1 mi N.W. of Sherburne Center. The phyllonites derived from schists and gneisses of the Mount Holly are not easily distinguished here from the schistose metagraywackes of the Tyson Formation,--a bad place for a beginner!

15.9 Stop 4: Just S. of Sherburne Center. Deformed polymictic conglomerates in lower part of Tyson Formation. Admire, but please do not destroy, the pebbles of blue opalescent quartz near the south end of the outcrop.

16.8 Stop 5: Unconformity at base of Tyson Formation is exposed on right near northeast base of outcrop. Note graded beds and onlap relations in basal part of Tyson. This outcrop should be preserved with care; hammering will not improve it in any way!

18.0 Stop 6: Carbonaceous, pyritic phyllites, dolomites and dolomitic quartzites in central part of Tyson Formation. These rocks underlie the upper, dolomite member of the Tyson which here controls the course of the headwaters of the Ottauquechee River. The conspicuous quartz vein is probably related to boudinage.

18.3 Enter Killington Peak 7 1/2 minute quadrangle.

18.4 Turn left on small side road and cross river.

18.6 Stop 7: Overhanging ledges E. of road are albitic schists at the base of the Hoosac Formation. The basal beds of the Hoosac contain abundant magnetite. Dolomite at top of Tyson Formation is exposed beneath overhang and contains lenticular masses of iron oxides that were once mined farther south in Plymouth as iron ore. These are thought to be a metamorphosed terra rosa and to indicate a period of subaerial erosion between the deposition of the Tyson and Hoosac Formations. The iron ores are probably correlative with those in northwestern Vermont and southern Quebec at the contact between the White Brook Dolomite and the overlying West Sutton Formation (Booth, 1950, p. 1146-7).

18.9 Outcrops of dolomite on left below overhanging ledges of albite schist.

19.2 Rejoin U.S. 4 and turn left, outcrops on right are quartzites in Hoosac Formation.

20.2 West Bridgewater, enter Plymouth 7 1/2 minute quadrangle. Outcrops in gravel pit to N.E. are carbonaceous phyllites in upper part of Hoosac Formation.

20.6 Stop 8: Green chloritoid phyllites of Pinney Hollow Formation, some are faintly purplish owing to hematite. Assemblage is quartz-muscovite-paragonite-chlorite-chloritoid±hematite. To east along base of bank are phyllites with the assemblage quartz-muscovite-chlorite-albite-garnet-biotite-magnetite-pyrite.

21.6 Stop 9: Carbonaceous, sulfidic phyllite and interbedded carbonaceous quartzites in lower part of Ottauquechee Formation. Some beds contain biotite and small garnets.

22.4 Carbonaceous schist and greenstone in Ottauquechee Formation.

22.8 Stop 10: Quartz-sericite-biotite-garnet schists of Ottauquechee Formation. Outcrop shows interbedding of carbonaceous and non-carbonaceous varieties.

24.2 Stop 11: Typical schist of Stowe Formation. Note abundant quartz lenses. These probably represent silica produced by metamorphic reactions. Outcrop is in garnet zone.

24.6	Enter Missisquoi Formation.
25.2	Stop 12: Carbonaceous schist and gray quartzites of Whetstone Hill Member of Missisquoi Formation.
25.8	Bridgewater Corners, turn right (S.) on Route 100A.
26.4	Stop 13: "Pinstripe" in Missisquoi Formation. This is the characteristic rock type of the Moretown Member although these particular outcrops are in the Whetstone Hill Member. Quartz-garnet-magnetite layers are probably recrystallized Fe-Mn cherts. Note rosettes of grunerite, but please spare for subsequent field trips.
27.1	Road bears right. Next four miles up Pinney Hollow repeats the Ottauquechee River section in reverse order.
27.2	Re-enter Stowe Formation.
27.9	Enter Ottauquechee Formation.
28.9	Carbonaceous schists and quartzites, Ottauquechee Formation.
29.2	Enter Pinney Hollow Formation.
30.6	Green chloritoid phyllites of lower Pinney Hollow Formation. Some purplish bands with hematite.
30.9	Enter Hoosac Formation.
31.0	Dolomites on left in upper part of Hoosac Formation.
31.4	Bear right to Plymouth Village.
31.6	Center of town, proceed straight ahead by cheese factory.
32.1	Bear left at fork in road (old limekilns on left).
32.3	Stop 14: Dolomite breccia (west of road) in Plymouth Member of Hoosac Formation. This rock closely resembles the dolomite breccia in the basal beds of the Rutland Dolomite at Stop 1. Outcrops in woods west of pasture are of underlying quartzites resembling the Cheshire Quartzite at Stop 1. The same quartzites may be seen 0.7 miles S. on Route 100A at Plymouth Notch. Descent from Plymouth Notch to Plymouth Union gives a fairly complete section through albitic schists of lower part of Hoosac Formation. Contact with dolomite of Tyson Formation is exposed just N.E. of junction with Route 100.

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Trip B-11

GEOLOGY OF THE GUILFORD DOME AREA, SOUTHEASTERN VERMONT

by

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Introduction

The Guilford dome lies within the broad outlines of the regional Connecticut Valley-Gaspé synclinorium. This synclinorium, principally underlain by Siluro-Devonian rocks, separates the Oliverian gneiss-cored domes of the Bronson Hill anticlinorium to the east from the Green Mountain anticlinorium to the west. The Guilford dome is part of a belt of domes that extends southward from east-central Vermont to Connecticut, west of the Connecticut River, analogous to but more widely spaced than the domes of the Bronson Hill anticlinorium. Large recumbent folds are found in the strata mantling these domes in eastern Vermont (Doll et al., 1961; Rosenfeld, 1968). The Standing Pond Volcanics is an important marker unit outlining many of these recumbent folds and domes. The axial surfaces of the recumbent folds have been arched by the later doming. The arcuate, closed, double band of the Standing Pond Volcanics around the southern end of the Guilford dome (Fig. 1) outlines such a refolded recumbent fold. One of the main purposes of the field trip is to investigate this fold and the proposed east-facing recumbent anticline above it. Other stops will be made to view the Black Mountain Granite, an important key in determining the time of deformation; the Siluro-Devonian Waits River Formation in the exposed core of the dome; and the Putney Volcanics, which separates the "Vermont" and "New Hampshire" sequences.

Acknowledgements

Geological mapping of the Guilford dome area was part of a Ph.D. thesis at Harvard University under the direction of Professors M. P. Billings and James B. Thompson, Jr., whose

help the author would particularly like to acknowledge. I would also like to thank the many persons who assisted during the course of the field work. Financial assistance of the Reginald and Louise Daly Fund, Harvard University, is gratefully acknowledged.

Stratigraphy

Please refer to Skehan and Hepburn (this volume), Stratigraphy of the East Limb of the Green Mountain Anticlinorium, Southern Vermont, for a brief description of most of the stratigraphic units and for a regional correlation chart. The units most pertinent to this trip are summarized below.

Middle Ordovician

BARNARD VOLCANIC MEMBER, MISSISQUOI FORMATION: 4000-8000 feet thick. Massive porphyritic and non-porphyritic amphibolites, feldspar-rich gneisses, and layered gneisses.

Siluro-Devonian

SHAW MOUNTAIN FORMATION: 0-20 feet thick. Quartzite and quartz-pebble conglomerate, hornblende fasciculite schist, amphibolite, and mica schist.

NORTHFIELD FORMATION: 1000-2500 feet thick. Gray mica schist with abundant almandine porphyroblasts, minor impure quartzite and impure punky-brown weathering marble.

WAITS RIVER FORMATION: 3000-7500 feet thick. Mica schist (phyllite at lower metamorphic grades) and calcareous mica schist with abundant interbeds of punky-brown weathering, impure marble; thin interbeds of micaceous quartzite. Quartzitic member: feldspathic and micaceous quartzite interlayered with muscovite schist.

STANDING POND VOLCANICS: 0-500 feet thick. Medium-grained amphibolite and epidote amphibolite; garnet-hornblende fasciculite schist. Eastern band: plagioclase-biotite-hornblende-quartz granulite and gneiss.

GILE MOUNTAIN FORMATION: 2500-5000 feet thick. Light gray to gray, micaceous and feldspathic quartzite and mica schist; gray fine-grained phyllite and slate with interbedded, thin micaceous quartzite; and rare impure marble. Marble member: black phyllite with interbeds of punky-brown weathering, impure marble and micaceous quartzite.

PUTNEY VOLCANICS: 0-400 feet thick. Light, greenish gray phyllite; buff to light brown weathering feldspathic phyllite; thin beds of feldspathic granulite; and minor gray slate. Conglomeratic member: lenses of polymict conglomerate with a gray slate matrix; pebbles abundant to scarce.

LITTLETON FORMATION: 5000-6000 feet thick. Gray slate or phyllite with interbedded quartzite.

Early to Middle Devonian Intrusive Rocks.

BLACK MOUNTAIN GRANITE: Medium-grained two-mica granodiorite, correlated with the New Hampshire Plutonic Series (Billings, 1956).

No new definitive evidence for the facing of the Waits River, Standing Pond, and Gile Mountain Formations has yet been found by the author. However the sequence, oldest to youngest, of Waits River, Standing Pond, and Gile Mountain, as shown on Figure 1 is favored, although a possible inversion of this order cannot be ruled out.

The Putney Volcanics (Stops 1 and 2) consists of a belt of rocks that were formerly included in the Standing Pond Volcanics (Doll et al., 1961; Trask, 1964). Since the proper correlation of these rocks has not yet been established, Hepburn (1972) designated them as a separate formation.

Structural Geology

The major tectonic features in the Guilford dome area formed during the Acadian orogeny, between the end of sedimentation in the Early Devonian and the crystallization of late, unoriented, coarse muscovite crystals in the Black Mountain Granite 377-383 m.y. ago (Naylor, 1971). Late normal faulting and possibly some minor folding occurred during the Triassic. The two major stages of deformation in the area include (1) the development of large recumbent folds, followed by (2) the rise of the Guilford dome.

The doubly-closed loop of the Standing Pond Volcanics around the southern part of the Guilford dome outlines the Prospect Hill recumbent fold, named for exposures at the hinge (Stop 3). The Gile Mountain Formation forms the core of the fold. Originally the Prospect Hill fold had a subhorizontal

axial surface and a hinge striking northeast-southwest. The subsequent doming about a roughly N-S axis arched the axial surface of the recumbent fold, so that now the hinge plunges moderately northeast and southwest away from the axial trace of the Guilford dome. An early, tight, now overturned, steeply east-dipping synform must lie between the Standing Pond bands in the doubly-closed loop and a third band lying to the east of the Guilford dome (Fig. 1). The hinge line where the Standing Pond rocks cross the axial surface of this synform is not seen in the Brattleboro area and is presumably buried. This synform, the Northfield Formation around the north end of the Guilford dome, and the Fall Brook anticline which exposes the Barnard Volcanics, are interpreted as the upper (anticlinal) portion of the Prospect Hill fold (Fig. 1, Cross-section A).

It is very likely that the Prospect Hill fold is continuous with the Ascutney sigmoid in the Saxtons River quadrangle to the north (Rosenfeld, 1968; Doll et al., 1961). If this is true, the hinge of the Prospect Hill fold must turn more northerly a short distance north of Stop 3.

The Guilford dome, which occupies much of the central portion of the Brattleboro quadrangle (Fig. 1), is a large, elliptical, doubly-plunging anticline formed during the second major stage of deformation. The Waits River Formation forms the exposed core of the dome. The foliation dips away in all directions from the axial trace, which strikes slightly east of north and plunges moderately to the north and south at the ends of the anticline. The axial surface of the dome dips very steeply to the west. A small depression in the exposed central portion of the dome divides it into a northern and southern lobe. The axial trace of the dome is closer to its eastern side. Here, the foliation has steep dips a short distance east of the axial trace. Dips are more gentle to the west. Bedding with a schistosity parallel to it has been arched by the dome.

It is likely that the two major stages of deformation were not greatly separated in time.

Minor Folds

Minor folds of at least five different stages are present in the Guilford dome area and the Brattleboro syncline to the east of the dome. These stages of minor folding are summarized below:

F1. Small isoclinal folds in layering, with schistosity developed parallel to the axial surfaces (Stop 3).

F2. Tight to isoclinal folds congruous with the large-scale recumbent folding (Prospect Hill fold). These fold the schistosity and the F1 folds. Weak to moderate axial-planar cleavage. Plunge moderately NE. or SW.

F3. Open folds, particularly west and south of the Guilford dome. Excellent slip-cleavage developed parallel to the axial surfaces. The axial surfaces generally strike NE. and dip steeply NW. The hinges plunge moderately NE. Excellent crinkle lineations occur at the intersection of this slip-cleavage and the schistosity surfaces in the pelitic rocks.

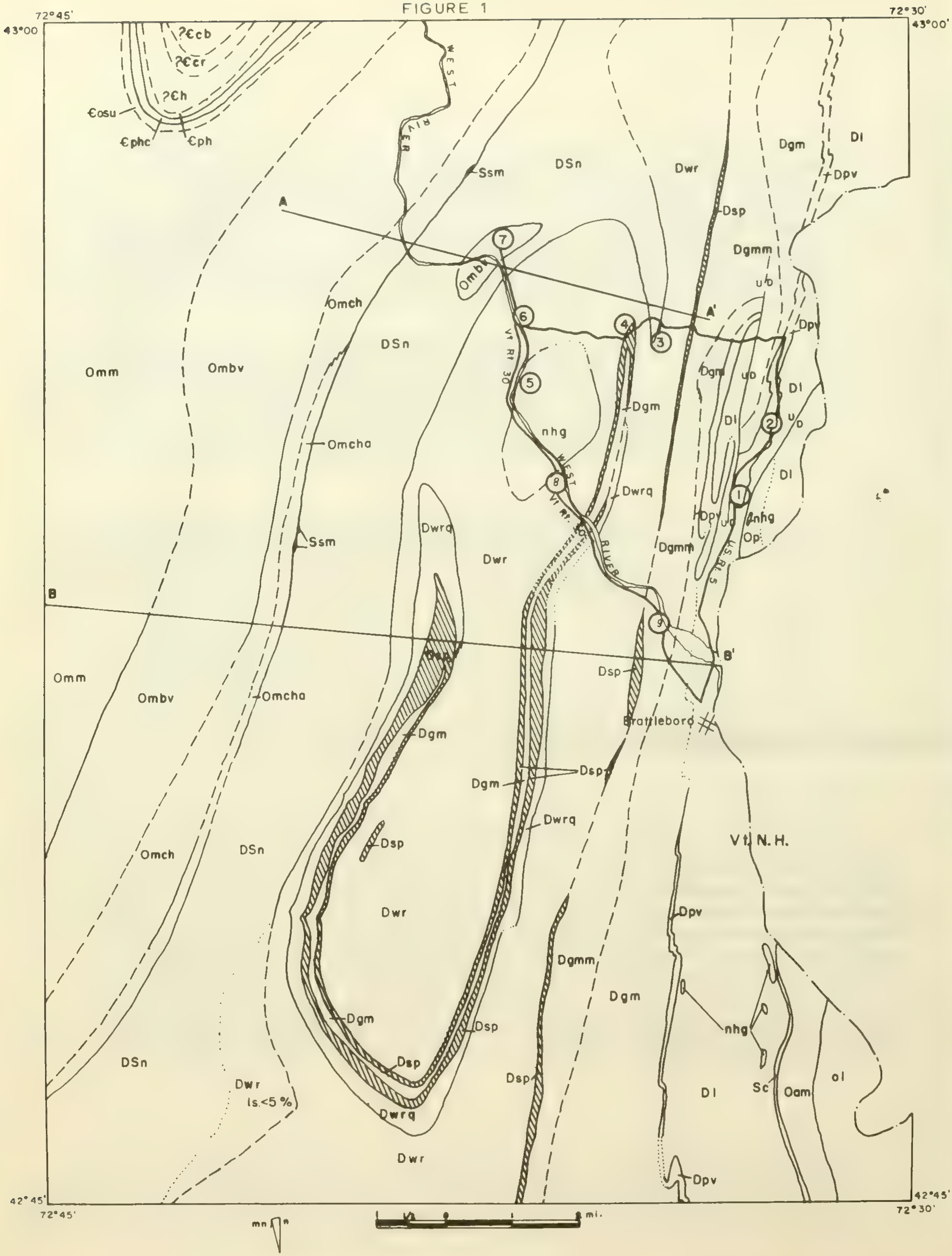
F4. Open folds, buckles or warps in the foliation that are of one or more generations and fold the slip-cleavage.

F5. Large open folds found only in the eastern part of the area (Fig. 1) that offset the Putney Volcanics with an east-side-north movement. Plunge is moderately to steeply north. Kink bands also found along the eastern part of Figure 1 are the youngest minor folds and may be related to the above F5 folds or may be younger.

Metamorphism

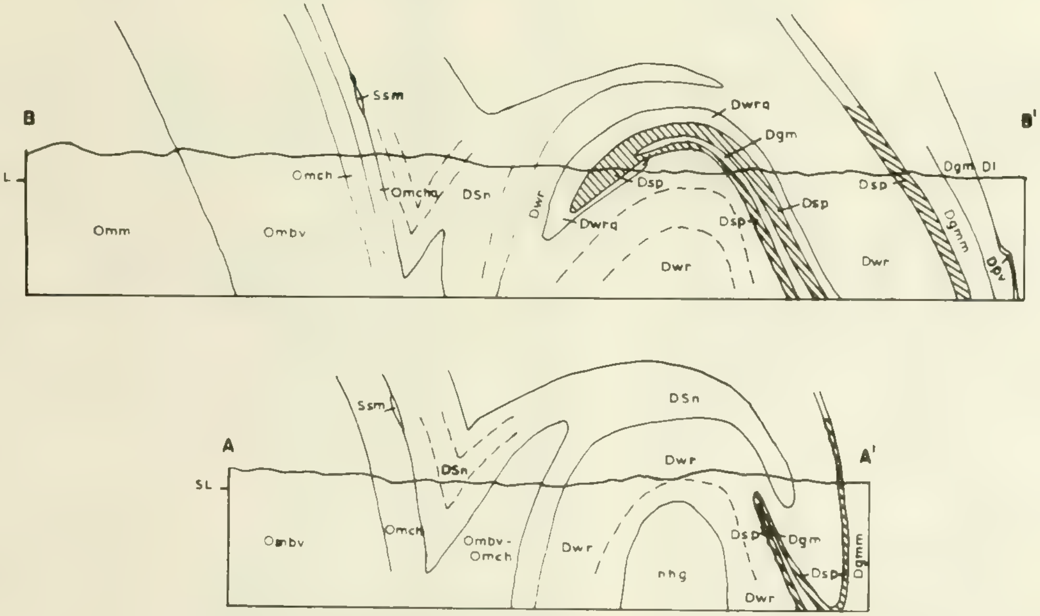
A belt of low-grade metamorphic rocks (chlorite zone) occurs in the eastern part of the area and roughly follows the Connecticut River. This low is of regional extent (Thompson and Norton, 1968) and separates terrains of higher metamorphic grade along the Bronson Hill anticlinorium from those in the domes of eastern Vermont. The highest grade of regional metamorphism in the Guilford dome area, staurolite-kyanite zone, is centered on the dome. The peak of metamorphism probably closely followed the doming stage of major deformation. During the earlier recumbent folding, the grade of metamorphism did not exceed the garnet zone.

FIGURE 1



GEOLOGIC MAP OF THE BRATTLEBORO AREA

FIGURE 1



LEGEND

Dgm	Gile Mountain fm		
Dgmm	Gile Mountain marble mem.		
(hatched)	Standing Pond vol	Dpv	Putney vol
Dwrq	Waits River qtzt. mem.	Dl	Littleton fm
Dwr	Waits River fm		
DSn	Northfield fm.		
Ssm	Shaw Mountain fm	Sc	Clough quartzite
Omcha	Missisquoi fm. Cram Hill amph.		
Omch	Missisquoi fm Cram Hill mem		
Ombv	Missisquoi fm Barnard vol	Op	Partridge fm.
Omm	Missisquoi fm. Moretown mem.	Oam	Ammonoosuc vol.
Cosu	Ottauquechee - Stowe fm		
Cphc	Chester amph.		
Cph	Pinney Hollow fm	nhg	New Hampshire Plutonics
?Ch	Hoosac fm.		
?Ccr	Cavendish fm. Readsboro mem		
?Ccb	Cavendish fm. Bull Hill mem.		
ol	Oliverian Plutonics		

contact

approximate or gradational contact

inferred contact

Geology of easternmost part modified after TRASK 1964

3 Field Trip Stops

Geology of the Guilford dome area, southeastern Vermont.

Road Log for Field Trip, Sunday Oct. 15
J. Christopher Hepburn, Leader

Assemble at STOP 1 at 10:30 a.m. This will allow plenty of time for participants leaving Burlington by 8:00 a.m. to arrive. The trip will never be more than a few miles from I91 for those who must leave early. Bring lunches.

Topographic map: Scheduled stops will be in the Brattleboro 15 minute quadrangle, Vermont-New Hampshire. The Geologic Map of Vermont by Doll *et al*. (1961) may also be of interest and is available from the Vermont State Library, Montpelier for $4.00.

Mileage	
	From Burlington take I89 south to I91. Then I91 south to Exit #3, the first Brattleboro exit from the north, marked "To Route 9 east, Keene, N.H.; and Route 5, Brattleboro".
0.0	At the junction of Routes 5, 9, and 91 north of Brattleboro by Howard Johnson's Restaurant just off Interstate Exit #3, turn left (north) onto Route 5.
0.7	Overpass over I91.
0.9	Brattleboro-Dummerston town line.
1.2	STOP 1. Meeting Place, PUTNEY VOLCANICS. Park in rest and picnic area on the east side of Route 5. The Putney Volcanics (Hepburn, 1972) in this area consists of fine-grained, poorly foliated, light greenish gray quartz-plagioclase-muscovite phyllites and granulites with interbedded gray slates. The granulites and feldspathic phyllites weather buff to light brownish gray, characteristic of feldspar-rich rocks. Many of the foliation surfaces have a notable silky sheen. Small, brownish pits where carbonate has weathered out are common. The granulite beds may show a fine lamination. A few lenses of quartz-pebble conglomerate may be seen along Route 5 south of the highway pull-off but are much better developed at Stop 2. The rocks have been metamorphosed to the chlorite zone at this locality. Continue north on Route 5.
1.4	Outcrop of Putney Volcanics to the east.
1.5	Outcrop of Putney Volcanics to the west.
2.1	Slate quarry in Littleton Formation to the east.

2.3 STOP 2. PUTNEY VOLCANICS, CONGLOMERATIC MEMBER. Park at left (west) side of road in the highway pull-off.

Examine outcrops of gray slate in the Littleton Formation on the east side of Route 5. Then walk 0.1 mile north through woods to an abandoned chicken-yard beside houses to west of Route 5. Outcrops are of the conglomeratic member of the Putney Volcanics. The contact of this conglomerate with the Littleton Formation represents the division between the "Vermont" and "New Hampshire" sequences in this area. The conglomerate contains both quartzite and slate pebbles in a slate matrix. (As this is the best exposure and type locality for the conglomerate, NO HAMMERING--PLEASE!). The excess of matrix over clasts in the conglomerate indicates it best fits Pettijohn's (1957) classification as a paraconglomerate. Pettijohn (1957, pp. 265-266) states that "it now seems probable in light of our knowledge of turbidity currents and related mudstones that most of these abnormal conglomerates [the paraconglomerates] are the product of subaqueous mudslides or slurries".

A few small porphyroblasts of light pink garnet occur here. The outcrop is included in the chlorite zone, however, as probe analyses indicate these garnets contain up to 15.9 weight percent MnO. (The garnet isograd has been mapped on the first appearance of almandine in the pelitic rocks.)

Immediately west of the conglomerate in this outcrop, the Putney Volcanics consists of slate with feldspathic granulite interbeds up to 2 feet thick. The granulites have fine laminations. M. P. Billings (1971, personal communication) indicated that a number of years ago he had found cross-bedding in these granulites that indicated tops to the west. This stop has become more overgrown in recent years, since the chickens left.

West of the abandoned chicken-yard a sequence of phyllites and feldspathic granulites similar to those at Stop 1 is exposed on the side of the hill.

Return to cars. Continue north on Route 5.

2.4 Road junction with dirt road on right. Continue north on Route 5.

2.6 Roger's Construction Co. yard on right (east), possible alternate parking for Stop 2.

2.9 Dutton Pines State Forest.

3.4 Road junction with road to East Dummerston; continue on Route 5. Outcrop of Putney Volcanics to west.

3.8 Road junction. Turn left (west) on road to East Dummerston and Dummerston Center.

4.7 Road junction in East Dummerston; continue straight.

4.8 Junction with road on right; continue straight.

4.9 Outcrop of Waits River Formation.
5.9 Dummerston Center. Turn sharp left (south).
6.0 STOP 3. NORTHFIELD FORMATION. Park along side of road.
Walk west to outcrops of the Northfield Formation exposed near the hinge area of the recumbent anticline above the Prospect Hill fold (See Fig. 1). The Northfield here is a gray well-foliated mica schist with conspicuous garnet porphyroblasts and fewer porphyroblasts of biotite and staurolite. A few thin interbedded quartzites are also present.
Turn around; return north to Dummerston Center.
6.1 Dummerston Center. Turn left (west) on paved road past the fire station.
6.5 STOP 4. HINGE OF PROSPECT HILL FOLD, WAITS RIVER FORMATION AND STANDING POND VOLCANICS. Park in road pull-off on north side of the road just before the curve.
The Standing Pond Volcanics outline the northeasterly plunging hinge of the Prospect Hill recumbent fold at this locality (Fig. 1). A 1/2 mile traverse will be made around the hinge, following the contact between the amphibolites of the Standing Pond Volcanics and the schists, calcareous schists, and impure marbles of the Waits River Formation. This traverse presents an excellent opportunity to view a well-exposed hinge of a major recumbent fold. The contact is sharp and is easy to follow. The traverse starts just east of the pull-off near a very small creek along the eastern contact of the Standing Pond Volcanics. Follow this contact to the north and around the northeasterly plunging hinge of the recumbent fold, which closes on the lower south-facing slopes of Prospect Hill. Continue along the contact southward (now the western contact of the Standing Pond with the Waits River). The paved road is encountered again 1/4 mile west of the starting point.
If time permits, Prospect Hill will be climbed for the excellent view from the open summit (perhaps lunch). Please be particularly careful on this traverse with litter and the indiscriminate use of hammers. We are able to make this stop only with special permission.
Particular note should be made of the minor folds during the traverse. The most common folds are the F2 generation, those formed congruously with the recumbent folding. These plunge NE. and show a reversal in drag sense around the hinge. A few F1 minor folds that predate the recumbent folding, have the principal schistosity parallel to their axial surfaces, and are refolded by the F2 folds are visible in outcrops near the road.
Return to cars; proceed west on paved road.
6.7 Outcrops of the Standing Pond Volcanics in the hinge of the Prospect Hill recumbent fold.

6.8 Contact of the Standing Pond Volcanics with the Waits River Formation.

6.9 Junction with dirt road to south; continue straight on paved road.

7.4 Outcrop of aplitic dike associated with the Black Mountain Granite.

7.8 Junction with road from right (north); continue straight.

8.5 Road junction; take sharp left onto dirt road.

9.3 STOP 5. BLACK MOUNTAIN GRANITE. Park by abandoned quarry buildings and follow path east to the abandoned Presbury-Leland granite quarry.

The Black Mountain Granite is a late synorogenic to post-orogenic two-mica granodiorite correlated with the New Hampshire Plutonic Series (Billings, 1956). Note the weak foliation produced by the alignment of the fine-grained micas. Coarse, unoriented muscovites that are younger than this foliation have been dated by Naylor (1971) from this locality. He obtained Rb/Sr ages of 377 m.y. and 383 m.y. for these muscovites, which sets a minimum age for the pluton as late Early to early Middle Devonian.

West- to northwest-dipping sheeting is well exposed in the quarry walls. Note particularly the increased thickness of the individual sheets with depth.

STOP 5a.

Walk west from the quarry to the banks of the West River. The contact of the granite body with the surrounding Waits River Formation is well exposed here. Dikes and sills of granite and aplite are numerous within a few hundred feet of the contact and may indicate a stoping mechanism for the emplacement of the granite pluton. The dikes cross-cut bedding and the principal schistosity. Some have a weak foliation roughly parallel to the regional schistosity but clearly post-date the major deformation. The country rocks near the granite have been altered by contact metamorphism, in addition to being regionally metamorphosed to the staurolite-kyanite zone.

Return to cars; turn around and retrace route north to the main road.

10.1 Junction with paved road; continue straight (north).

10.2 STOP 6. WAITS RIVER FORMATION. Park just beyond the entrance to the covered bridge, heading north.

Outcrops typical of the Waits River Formation in the center of the Guilford dome are seen along the east bank of the West River. The rocks are interbedded impure marbles, calcareous mica schists, and mica schists. Most of the minor folds present here are assigned to the F2 stage and developed congruously with the large-scale recumbent folding. They were refolded into their present attitude by the rising of the Guilford dome.

Return to cars; proceed straight (north) on the dirt road along the east side of the West River.

10.7 Junction with road to right; continue straight.

11.2 STOP 7. BARNARD VOLCANICS. Park along the road above the east end of the old West Dummerston Dam. Climb down the steep bank (Use caution.) to the west end of the now abandoned dam.

The Middle Ordovician Barnard Volcanics are exposed here in the center of the Fall Brook anticline, which forms the core of the proposed recumbent anticline above the Prospect Hill recumbent fold (Fig. 1). At this stop the rocks include amphibolites and felsic gneisses. Minor amounts of rusty-weathering schist similar to the Cram Hill are present along with the Barnard in this anticline but have not been designated separately on Figure 1.

Turn around; retrace route south to the covered bridge.

12.2 Covered bridge; turn right; cross the bridge. At the west end, turn left (south) onto Route 30.

12.9 West Dummerston Village. Note Black Mountain and the granite quarry to the east across the West River.

13.3-13.6 Outcrops of the Waits River Formation.

13.8 Iron bridge to left; junction of road to the right. Continue straight on Route 30. Outcrops of granite in the brook to the west.

15.2 STOP 8. WAITS RIVER FORMATION ALTERED BY CONTACT METAMORPHISM. Park at the side of Route 30 by the large road-cut on the right (west).

The Waits River Formation in this outcrop is near the contact of the Black Mountain Granite. Calc-silicates (particularly actinolite and diopside) are well developed in the impure marble beds. Diopside has not been observed in the Waits River Formation of the Guilford dome area outside of the contact aureole of the Black Mountain Granite.

Continue south on Route 30.

16.8 Roadmetal quarry in the Waits River Formation to the west.

17.0 Outcrop of Waits River Formation.

17.7 STOP 9. GILE MOUNTAIN FORMATION, MARBLE MEMBER. Park at left in the pull-off under the I91 overpass.

Outcrops under the overpass are fairly fresh exposures of the marble member of the Gile Mountain Formation, metamorphosed to the biotite zone. The impure marble beds (already starting to obtain the distinctive punky-brown weathering rind) similar to those in the Waits River Formation are interbedded with phyllites. The percentage of micaceous quartzite beds

is fairly high here (approximately 15 percent), as is typical of this member.

END OF FIELD TRIP

Continue south 1.5 miles to Brattleboro for junctions with the major highways.

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Trip B-12

STRATIGRAPHIC AND STRUCTURAL PROBLEMS OF THE SOUTHERN PART OF THE GREEN MOUNTAIN ANTICLINORIUM, BENNINGTON-WILMINGTON, VERMONT

by

James W. Skehan, S.J.*

INTRODUCTION

This field trip is an introduction to several aspects of problems that have vexed students of the geology of the Green Mountains, the Berkshires and the Taconic Mountains for decades. Hitchcock very early (1861) noted that the rock units flanking the eastern side of the Precambrian core of the Green Mountains were different from those of its western flank (Fig. 1). Prindle and Knopf (1932) explained this and the juxtaposition of the two contrasting sequences by inferring the existence of the Hoosac Thrust which they and MacFadyen (1956) mapped as far north as Heartwellville. They also mapped the "Cambrian outliers" in the dominantly Precambrian terrain of the Green Mountain core (Figs. 1 and 2). Skehan (1961 and this paper) extended the Hoosac fault northeasterly and infers tentatively that it marks the trace of the plane of angular discordance between the Mt. Holly Complex and the Cavendish Formation. Dale (1914-16) was the first to map this same contact of the Green Mountain core, which he referred to the Algonkian, with the younger rocks (Cambrian) to the east in Searsburg (Stops 7 and 8). He regarded this boundary as an angular unconformity. The related problem of recognizing the source area and mechanism of emplacement of the Taconic allochthon has been addressed by many students of Green Mountain and Taconic geology.

Skehan (1953 and 1961) traced rock units mapped by Thompson (1950) and Rosenfeld (1954) in the Ludlow and Saxtons River quadrangles respectively through the Wilmington area to the Massachusetts border. Mapping in adjacent parts of Massachusetts has been carried out by Pumpelly, Wolff and Dale (1894), Osberg (1950), Chidester *et al*. (1951), Segerstrom (1956), Herz (1958), Hatch (1967) and Hatch, Stanley and Clark (1970) who have traced the units of the Vermont sequence south to Connecticut.

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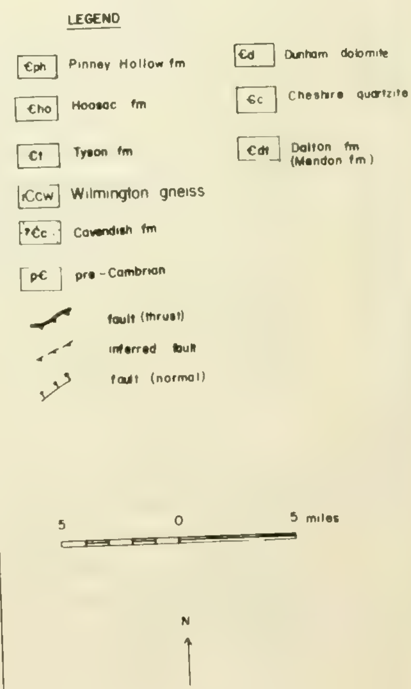
FIGURE 1

GEOLOGIC MAP SHOWING THE CAVENDISH FORMATION RELATIVE TO THE GREEN MOUNTAIN ANTICLINORIUM AND RELATED FORMATIONS



All of these workers recognized that the rocks of the Taconic Allochthon (Zen, 1967, Bird, 1969) are similar to those of the eugeosynclinal sequence east of the Green Mountains allowing for differences in the grade of metamorphism. Several of these geologists have research projects in progress which bear on a solution to problems of the present field trip.

The present field trip proposes to introduce the participants to representative rock types of the western Cambrian sequence (Stops 1 and 2) and its continuation on the eastern flank (Stop 12) as well as to the Precambrian core rocks of the Green Mountains (Stops 3, 4, 5, and 7). Additionally several of the component stratigraphic units as well as structural relationships of the questionable Cambrian sequence of the Cavendish Formation of Doll _et al_. (1961) (Stops 6, 8, and 11) to other units will be studied.

STRATIGRAPHY

The stratigraphic succession of the area of the field trip (Fig. 2) includes the Mt. Holly Complex of Precambrian age, the Cavendish Formation including the Wilmington Gneiss of questionable Cambrian age and the Dalton and Cheshire Formations of Lower Cambrian age and the Hoosac Formation of Cambrian age.

Mount Holly Complex

The Mt. Holly Complex (Skehan, 1961, pp. 28-45) forming the core of the Green Mountain Anticlinorium consists of several units:

Microcline Gneiss. The largest part of the Green Mountain core in the Wilmington-Woodford area is underlain by coarse-grained banded biotite-epidote-quartz-microcline augen gneiss. Commonly the quartz is blue. This unit is lithologically similar to and in many exposures texturally identical with rocks of the Stamford Granite Gneiss. Except that blue quartz is absent in the Wilmington Gneiss, it is otherwise indistinguishable from the microcline gneiss of the Mt. Holly Complex (Skehan, 1961, pp. 29-31) and the Bull Hill Gneiss of the Cavendish Formation of Doll _et al_. (1961).

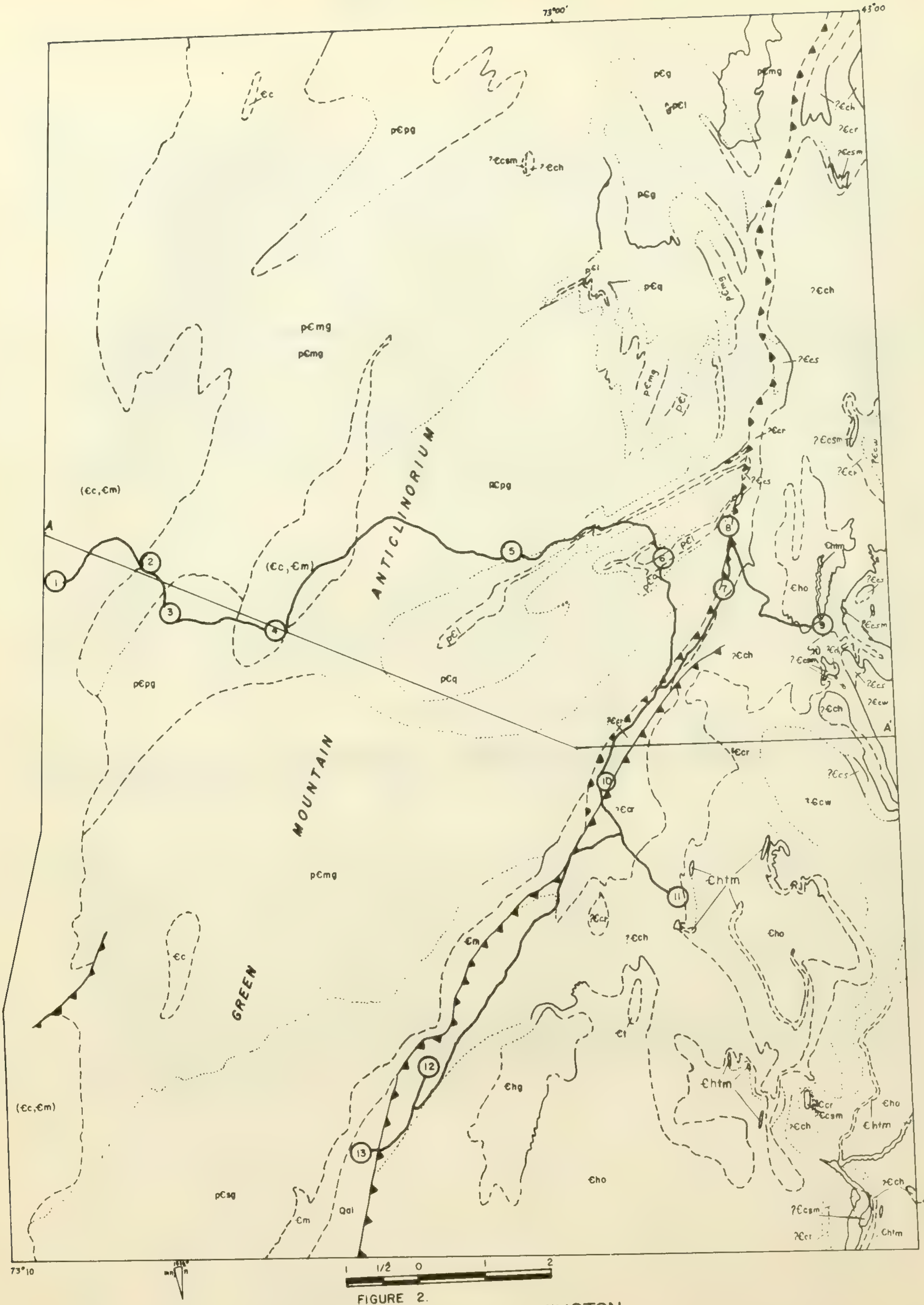


FIGURE 2.

GEOLOGIC MAP OF THE WILMINGTON WOODFORD AREA

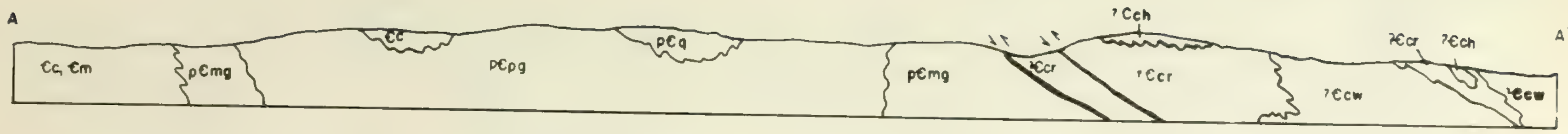


FIGURE 3

LEGEND

Qal	Quaternary alluvium		
€ho	Hoosac fm. (green schist = €hg)		
€c	Cheshire quartzite		
€m	Mendon fm. (Dalton of Doll et al., 1961)		
€htm	Turkey Mountain mem.		
€t	Tyson fm.		
?€ch	Heartwellville schist		contact exactly located
?€csm	Sherman marble		contact approximately located
?€cr	Readsboro schist		inferred or gradational contact
?€cs	Searsburg cong.		
?€cw	Wilmington gneiss		thrust fault
p€g	Plagioclase gneiss		
p€l	Lime silicates	7	stop
p€q	Quartzite		
p€pg	Plagioclase gneiss		
p€mg	Microcline gneiss		
p€sg	Stamford granite gneiss		

Cavendish Fm

Plagioclase Gneiss. (Harmon Hill Gneiss). Large areas of the Green Mountain core are underlain by dark, banded muscovite-biotite-epidote-plagioclase-quartz gneiss commonly containing lesser amounts of microcline and quartz in layers and pods, as well as beds of amphibolite (Skehan, 1961, pp. 31-35).

Stamford Granite Gneiss. This distinctive rock is a coarse-grained porphyritic gneiss with very large rectangular to rounded microcline crystals. The finer grained groundmass consists of blue quartz, albite, microcline, biotite, epidote and magnetite. This unit (Doll et al., 1961) is in many respects similar to the Bull Hill Gneiss of the Cavendish Formation. The Stamford Granite Gneiss is considered to be probably intrusive into the Microcline Gneiss unit (p€mg) and related rocks (Pumpelly et al., 1894).

Younger Metasedimentary Rocks. A distinctive sequence is developed in the eastern part of the Green Mountain core and consists of massive, buff to blue vitreous quartzite, blue quartz conglomerate; conglomeratic gneiss composed of angular to rounded microcline and granite gneiss pebbles; crystalline graphite-bearing, blue and white quartz-rich white gneiss; fine to very coarse-grained calc-silicate granulite, and blue and white quartz-plagioclase gneiss.

Cavendish Formation

Skehan (1961, pp. 46-65 and Pl. 1) mapped the following sequence in the area east of the Green Mountain core: the Searsburg Conglomerate Member, the Readsboro Schist unit, and the Sherman Marble Member of the Readsboro Formation. Additionally he mapped the Heartwellville Schist, which is lithologically similar to the Gassetts Schist of the Chester Dome, as a separate and younger unit. Doll et al. (1961) showed this sequence as the Cavendish Formation (Fig. 1) distinguishing the following units: the Sherman Marble, the Bull Hill Gneiss and the Readsboro-Gassetts Schist. In the present paper for the purposes of more general discussions we shall follow the usage of Doll et al. and use the term Cavendish to refer to this entire sequence of Searsburg-Heartwellville Schist.

It is useful, however, for detailed discussions of this particular area to further subdivide the Cavendish Formation of the Wilmington-Woodford area into its generally distinctive lithologies even though their stratigraphic position is not clear in all parts of the area (Fig. 2). In the present discussion the Wilmington Gneiss, lithologically similar to the Bull Hill Gneiss, is considered as closely related to the Cavendish Formation and is tentatively included in that sequence (Figs. 2 and 3).

Wilmington Gneiss

The Wilmington Gneiss named by Skehan (1961) is of uncertain stratigraphic position. It may be Precambrian in age, resembling as it does the microcline gneiss sequence of the Mt. Holly Complex of the Green Mountain core. On the other hand the apparently conformable relationship immediately beneath the Hoosac and Tyson Formations along their eastern contact (Fig. 1) suggests strongly the possibility that the Wilmington Gneiss may be of Cambrian age. The complex and very complicated relationships of the Wilmington Gneiss to the members of the Cavendish Formation of Doll et al. (1961) along the western contact makes a decision as to the age of the Wilmington Gneiss impossible at this time.

The Wilmington Gneiss consists of a medium to very coarse-grained, well-banded, somewhat foliated biotite-epidote-quartz-microcline-augen gneiss. The microcline is gray to pink and occurs as lenticular augen and flaser in which the average long diameter is about 7 mm. Locally the augen may reach 8 inches in length and are usually flattened into the plane of the foliation. Quartz rods and linearly aligned streaks of biotite are a common feature of the Wilmington Gneiss.

The Wilmington Gneiss may be the correlative of the Bull Hill Gneiss of Doll et al. (1961) an exposure of which is only one mile north of and on line with the northernmost exposure of the Wilmington Gneiss of the Wilmington quadrangle (Skehan, 1961, Pl. I).

Searsburg Conglomerate Member. The Searsburg Conglomerate Member is typically a blue or white quartz, albite and/or microcline-pebble conglomerate in a dark biotite-muscovite-carbonate-albite-quartz schist matrix. Thin bedded vitreous buff, white and gray quartzite in dark mica quartz schist is closely associated with the conglomeratic facies.

Readsboro Schist. The Readsboro Schist as presently understood by the writer is indistinguishable in hand specimen or outcrop from the Hoosac Formation consisting as it does of gray, brown and black, medium to coarse-grained muscovite-biotite-albite-quartz schist locally containing variable amounts of chlorite, muscovite, chloritoid, paragonite and garnet. Albite megacrysts 2-15 mm. in diameter are characteristic of the formation. The Readsboro Schist encloses calcite and dolomite marble of the Sherman Member whereas no marble beds have so far been recognized in the Hoosac Formation. The Hoosac Formation does, however, contain amphibolite beds of volcanic origin. These two formations are thus mapped on the basis of these differences.

Sherman Marble Member. The Sherman Marble is a coarse to very coarse-grained white, mottled green and gray to pink, quartz-calcite marble with coarse crystals of graphite up to 1 cm. in diameter; actinolite or diopside-phlogopite-talc calc-silicate granulite; and fine-grained quartz-dolomite marble. This marble is more commonly enclosed in the albite schist sequence but in the northern part of Mount Snow (Pisgah) it occurs in the Heartwellville beds.

Heartwellville Schist. The Heartwellville Schist is the lithologic and possibly the stratigraphic equivalent of the Gassetts Schist of Doll *et al*. (1961) of the Cavendish Formation. In the Wilmington-Woodford area the lower part of the Heartwellville Schist consists dominantly of green chlorite-muscovite-(paragonite-chloritoid)-garnet-quartz schist whereas the upper part is dominantly coaly-black, rusty weathering muscovite-chlorite-garnet-quartz schist. In hand specimen or in outcrop these rocks are indistinguishable from their counterparts in the Pinney Hollow and Ottauquechee Formations except that the Heartwellville is characteristically more highly deformed.

Dalton Formation

The Dalton Formation of the Wilmington-Woodford area is separated from the overlying rocks of the Cavendish Formation on the southeastern flank of the Green Mountain Anticlinorium by the Hoosac Thrust and from the Mt. Holly Complex by an angular unconformity. The Dalton consists of thin-bedded schistose muscovite-blue quartz quartzite; biotite-albite-quartz schist; black chloritoid-muscovite-quartz phyllite (Mendon Formation of MacFadyen, 1956 and Skehan, 1961); and microcline-quartz gneiss. The Dalton Formation is of Lower Cambrian age since Walcott (1888) found fragments of *Olenellus* about 100 feet above the Stamford Gneiss contact near North Adams, Massachusetts in a quartzitic graywacke stratigraphically beneath a band of black phyllite considered to be the equivalent of the Moosalamoo and Mendon Formations.

Cheshire Quartzite

The Cheshire Quartzite is stratigraphically above the Dalton into which it grades. It is a buff, gray to light pink vitreous quartzite consisting of rounded quartz grains commonly showing overgrowths of quartz and cemented together by quartz and/or calcite. In many occurrences, the Cheshire shows primary sedimentary structures and is generally a ridge-former because of its resistance to erosion.

HOOSAC FORMATION

The Hoosac Formation (Hoosac Schist of Pumpelly et al., 1894) consists of gray, brown and black, medium to coarse-grained muscovite-biotite-albite-quartz schists locally containing variable amounts of chlorite, muscovite, paragonite and garnet. Rocks containing appreciable garnet commonly weather to a mottled rusty color. Albite megacrysts 2-15 mm. in diameter are characteristic of the formation, which is distinguished from the overlying Pinney Hollow Formation by the presence of more abundant albite megacrysts, its color, and its generally coarser and more granular texture.

The Turkey Mountain Member of the Hoosac Formation (named by Rosenfeld, 1954) is typically a dense dark green to black amphibolite commonly characterized by rounded to sub-angular white, gray, green or dark brown "amygdules" composed of quartz and albite commonly with included epidote, hornblende and garnet.

STRUCTURAL GEOLOGY

The area of the field trip is the southernmost part of the Green Mountain Anticlinorium which plunges south beneath the Cambro-Ordovician arenaceous and carbonate sequence of the North Adams-Williamstown area. The Cambrian beds of the western flank are overturned and in part faulted along high angle reverse faults (Fig. 2).

The Cambrian rocks of the southeastern flank of the Green Mountains are truncated by the easterly dipping Hoosac Thrust (Fig. 1). Rocks of the Cavendish Formation lie above the Hoosac Thrust and/or the Precambrian-Cambrian unconformity along the eastern Green Mountain front. This boundary between the Cavendish Formation and the Mt. Holly Complex is now considered tentatively by the writer to be a thrust fault since in this region the Precambrian beds show a strong angular relationship to the Cavendish beds (Fig. 2). Elsewhere in the Green Mountains where the Cavendish or the Tyson Formations contact the Precambrian rocks, beds on both sides of the contact have been rotated or smeared out by tectonic forces into apparent conformability adjacent to the boundary. At some distance from the contact, however, the angular difference is observable. The presence of strong angular discordance close to the contact of the Mt. Holly with the Cavendish Formation suggests that the Precambrian units have been truncated by thrusting.

The data presently available allow the following alternative interpretations:

(1) The Cavendish Formation, including the Wilmington Gneiss, is of Precambrian age; (2) the Cavendish including the Wilmington Gneiss is of Cambrian age but older than the Hoosac Formation of known Cambrian age and (3) the Cavendish and the Hoosac Formations are both of Cambrian age and are coeval facies of each other but the Hoosac now bears a thrust or some other complex structural relationship to the Cavendish. Skehan in 1961 offered the first alternative as his preferred interpretation at that time. Recognizing that each of these hypothesis are possible, his present understanding of the problem leads him now to prefer the second or third hypotheses with (3) being favored, although not proven, because it helps to explain more satisfactorily our present understanding of the relationship of the Cavendish to the Dalton Formation of the southeastern margin of the Green Mountain core as well as to the core rocks themselves (Figs. 1 and 2).

The fact that the Hoosac Formation (Fig. 3) overlies the rocks of the Cavendish Formation with an angular discordance led Skehan (1961) to consider these rocks of questionable Precambrian age and Doll *et al.* (1961) to regard them as of questionable Cambrian age.

TRIP LOG

Bennington may be reached by travelling south from Burlington on Route 7 (the shortest distance) or on I-91 (a faster highway) to Brattleboro and driving about 35 miles west on Route 9.

The primary references for this trip are:

Skehan, J.W., S.J., *The Green Mountain Anticlinorium in the Vicinity of Wilmington and Woodford, Vermont*: Bull. 17, Vermont Geological Survey, 159 p., 1961 ($3.00).

----, *Geologic Map of the Wilmington-Woodford, Vermont Area, from Bull. 17, Vermont Geological Survey, 1961 (25¢).*

Doll *et al.*, *Centennial Geologic Map of Vermont*, October, 1961 ($4.00).

(These three reference materials may be obtained from the State of Vermont, Department of Libraries, Montpelier, Vermont by enclosing remittance with order.)

NOTE: Proceed on your own to Stop 1 after which go to Stop 2, where the group will meet at 10:00 a.m. for a traverse along City Stream.

Mileage

0.00 Woodford-Bennington township line on Route 9 east of Bennington Center about 3.5 miles.

0.20 Stop 1. CHESHIRE QUARTZITE

A few hundred feet east of the township boundary of Bennington and Woodford on Route 9. Park off the highway near Mountain Melody Motel and walk south to the outcrop on the west side of the highway. These beds of Lower Cambrian Cheshire Quartzite consist of vitreous, buff to light pink, cross-bedded quartzite gently folded in an open anticline plunging westerly at approximately 15°. This fold is closely related spatially to but disharmonic as regards the major syncline whose southwesterly plunging axial trace passes near Woodford Hollow.

As indicated by sedimentary cross-bedding, these beds are right side up. Hand specimen and thin section examination of the rock shows rounded grains of detrital quartz. The beds of the eastern limb of this syncline rapidly become more steeply dipping and are even inverted toward the northeast in the direction of the western margin of the Precambrian core of the Green Mountains (Fig. 3), as the Cheshire Quartzite beds to the west give way to the stratigraphically lower beds of the Dalton Formation.

Return to cars and drive east on Route 9.

0.40 Outcrops of Cheshire Quartzite in the brook on the east. Much of the western slope of Harmon Hill to the east is upheld by the resistant beds of the Cheshire and Dalton Formations.

1.30 Junction of the Long Trail and Appalachian Trail with Route 9.

1.70 Junction of Woodford Hollow Road on the north with Route 9.

2.00 Stop 2. DALTON (MENDON) FORMATION AND MT. HOLLY COMPLEX

Park cars off the highway near the place where the high-tension power line crosses Route 9. Make a traverse on foot along City Stream in a westerly direction. This stop is an introduction to the Dalton Formation and to some of the Precambrian rocks and is designed to illustrate the problem of mapping the precise location of the Precambrian-Cambrian contact especially where the rocks on either side have been smeared into apparent conformability. Commonly, however, retrograde metamorphic effects in

Mileage (cont'd)

Precambrian rocks of appropriate composition are recognizable especially in thin sections. Moreover, many of the beds of the Lower Cambrian Dalton Formation, especially those consisting of vitreous quartzite containing rounded blue quartz sand grains and pebbles, are sufficiently distinctive to be recognized. The Dalton Formation additionally contains biotite-albite-quartz schist, and schistose muscovite-chlorite quartzite. In places, however, where biotite-plagioclase gneiss and microcline gneiss of the Dalton Formation overlies rocks of similar composition of the Mt. Holly Complex, from which they were derived by erosion, the precise location of the contact may be difficult to determine.

The Precambrian-Cambrian contact at this locality, about 350 feet west of the high-tension utility line, is placed at the western margin of a pyrite-bearing biotite-microcline gneiss which is closely associated with a chlorite-epidote amphibolite bed. The contact is considered to be folded or faulted since the rocks just mentioned are separated by a band of blue quartz conglomerate of the Dalton Formation from pink microcline gneiss to the east assigned to the Mt. Holly Complex (Skehan, 1961).

Proceed east on Route 9.

2.95 Pull off the highway at the large roadcuts near Dunville Hollow.

Stop 3. MOUNT HOLLY COMPLEX

Large roadcuts on both sides of Route 9 expose tight isoclinally folded bands of the dominantly plagioclase gneiss sequence of the Mt. Holly Complex of Precambrian age (Skehan, 1961, pp. 28-35). A less important component of the sequence here consists of microcline-rich bands and thin meta-amphibolites. The northeasterly trending well-developed folds are characterized by nearly vertical to steep westerly dipping axial planes. Post-metamorphic faults and shears, although variously oriented, are commonly developed essentially parallel to the axial planes of the folds (Skehan, 1961, Fig. 6). The second of two localities in the Wilmington-Woodford area where an unmetamorphosed basalt dike, considered to be of Triassic or Jurassic age, has been recognized is at this series of outcrops.

The rocks of the core of the Green Mountain Anticlinorium have been affected by both Precambrian and Paleozoic

Mileage (cont'd)

regional metamorphism. Broughton et al. (1962) refer the Precambrian metamorphism of the nearby rocks of the eastern Adirondacks to a "hypersthene zone" corresponding in its mineral assemblages to the higher grade part of the sillimanite-K feldspar zone as developed in the Paleozoic rocks of New England (Thompson and Norton, 1968). The rock sequence of the Mt. Holly Complex as mapped in the Wilmington-Woodford area bears a striking resemblance to that of the eastern Adirondacks, due allowance being made for the fact that the rocks of the Green Mountain Massif have been altered by retrograde Paleozoic metamorphism of approximately the biotite and garnet zones.

The dominantly dark biotite-plagioclase gneisses dip steeply to the west. Deformed pink microcline pegmatite layers and light gray feldspathic bands reveal that the sequence has been subjected to considerable deformation by being isoclinally folded. There are many bedding plane faults which are recognized as being essentially axial plane faults since the beds are so tightly folded.

Proceed east on Route 9 up the western flank of the Green Mountain Anticlinorium.

3.50 Large roadcut on the left in dark plagioclase gneiss is crosscut by folded Precambrian pegmatite.

3.85 On the right is a sequence of dark migmatitic gneisses. The migmatite is of microcline granite and pegmatite. Approximate western contact of the Cambrian beds of the Woodford "outlier" with the Mt. Holly Complex. Dark phyllite is well exposed in City Stream on the south side of Route 9 between here and Stop 4.

4.50 Black chloritoid-sericite-quartz phyllite of the Lower Cambrian Mendon Formation (MacFadyen, 1956; Skehan, 1961; and mapped as Dalton Formation by Doll et al., 1961) in City Stream on the south side of Route 9.

4.70 Stop 4. DALTON (MENDON) FORMATION

Park on the north side of the highway. Cross the road and examine the fine-grained chloritoid phyllite in the outcrops on City Stream.

Mileage (cont'd)

There are several localities in the core of the Green Mountain Anticlinorium where isolated outcrops of Lower Cambrian rocks of the Dalton Formation (Doll et al., 1961) are exposed of which the Woodford "outlier" is the largest exposure. It is about 5 miles long and 1 mile wide over much of its length. The northeasterly trending Woodford syncline is comprised of two major rock units: (1) the black carbonaceous biotite-sericite-chloritoid phyllite of the Dalton Formation and (2) the vitreous gray quartzite and schistose quartzite which may represent quartzite beds in the Dalton (Mendon) Formation.

The fact that the sequence of the Woodford syncline is comprised in large part of dark arenaceous phyllite suggests that its environment of deposition was more closely related to that of the Lower Cambrian Moosalamoo Phyllite (Doll et al., 1961) than to that of the dominantly arenaceous rocks which typify the Dalton Formation. Both are considered to be essentially of equivalent age.

These outcrops at Woodford are about 10 miles north of the locality near North Adams, Mass., at which Walcott (1888) found fragments of Olenellus mentioned above.

At this stop note that cleavage to bedding relationships are well developed. Cleavage chiefly dips more steeply than the bedding. Southeasterly dipping beds reveal that the structural analysis,however, fits no simple model of a typical synclinal structure developed by compression. Although various aspects of the Woodford "outlier" have been described by Prindle and Knopf (1932), MacFadyen (1956) and Skehan (1961) it is not definitely known whether these rocks are in normal depositional or in a thrust relationship to the underlying Precambrian Mt. Holly Complex.

Return to cars and proceed northeasterly on Route 9.

5.00 Near the entrance to the Prospect Mountain Ski area on the right, thin bedded, gray northeasterly-dipping sericite quartzite beds were exposed in 1959.

5.35 In Woodford Center near the church on the east side of the road folded, gray to black phyllite is exposed, the beds having the attitude, N. 75°W., 20°NE.

Mileage (cont'd)

5.75 200 feet southwest of the Peter Pan Motel on the left folded, thin-bedded quartzite beds crop out having the attitude, N.85°W., 35°SW. The folds, displaying a left-handed pattern (Skehan, 1961, p. 112sq.) plunge S.35°W. at 30°.

6.20 Big Pond on the left.

7.00 The divide at the crest of the Green Mountains intercepts Route 9 approximately at this location. Proceed downslope to the east. The topographic relief of the crest of the Green Mountains is generally subdued, outcrops are sparse and the swamp and forest cover are heavy. This condition, which is typical of large tracts in the Precambrian core of the Green Mountains, renders geologic mapping sufficiently difficult to impede detailed mapping and consequently a sophisticated understanding of the geology of the core of this massif.

8.25 Ann Marie's Restaurant -- the only all-weather restaurant between Bennington and Wilmington with the possible exception of motel-related dining facilities.

9.25 Stop 5. VIEW AND PICTURE STOP

Park on the north side of the highway at an abandoned gasoline station and cabins. To the north is a panoramic view of the breadth of the Green Mountain Massif with one of its highest peaks, Stratton Mountain in the Londonderry Quadrangle, visible in the distance. The rocks of the Mt. Holly Complex lie to the east of the Dalton Formation and Cheshire Quartzite, the ridge-formers on the near skyline to the northwest. In the far distance to the northwest may be seen Mt. Equinox of the Taconic Allochthon. To the east and northeast is the very prominent Mt. Snow (Pisgah)-Mt. Haystack Ridge comprised of questionable Cambrian metasediments of the Cavendish Formation of Doll *et al.* (1961).

Return to cars and proceed east on Route 9.

11.40 Junction of Route 9 with Route 8. Proceed south on Route 8.

11.95 Stop 6. VIEW AND PHOTO STOP

Park off the road and out of the line of traffic. Rusty weathering calc-silicate granulites are exposed in small road outcrops. This stop is near the eastern margin of the Precambrian core of the Green Mountains, which is

Mileage (cont'd)

bounded on the east by the easterly dipping Hoosac Thrust. The intensely deformed rocks of the Cavendish Formation rise up in the Haystack Mountain and Mount Snow (Pisgah) ridge. Their higher slopes are typically capped by the resistant dark muscovite-garnet-chlorite-quartz schists (Heartwellville Schist of Skehan (1961) and Gassetts Schists of Doll et al. (1961)). The Harriman Reservoir, filling a former river valley in the Wilmington Gneiss, may be seen to the east-southeast as viewed along the valley occupied by the east branch of the Deerfield River. Hogback Mountain on the distant skyline is held up by the Pinney Hollow garnet-muscovite-quartz schists and the Chester Amphibolite, the Ottauquechee and Stowe Formations, the schistose portions of these units being nearly identical in composition to rocks of the Cavendish Formation.

Proceed south on Route 8.

13.95 Junction of Route 8 with Sleepy Hollow Road. Farrington Cemetery is on the southeast corner of the junction. Turn left on Sleepy Hollow Road, and proceed 2 miles northeasterly to Bond Brook. Park off the road as best you can.

15.95 Stop 7. READSBORO AND HEARTWELLVILLE SCHISTS

Proceed on foot in an easterly direction along the north side of the swampy area. The stratigraphic section in Bond Brook consists of biotite-muscovite-garnet-albite-quartz schists overlain by garnetiferous chlorite-muscovite-quartz schist of the Cavendish Formation, these being identical in lithology with the Hoosac and Pinney Hollow Formations. The main thrust (and/or unconformity) is probably just west of Sleepy Hollow Road at this locality. Return to cars and proceed northerly toward Route 9.

16.35 Bridge over the penstock aqueduct which carries water from Searsburg Dam to Medburyville Power Plant.

16.55 Junction of Sleepy Hollow Road with Route 9. The trace of the boundary between the Mt. Holly Complex and Cavendish units (the Algonkian-Cambrian boundary of Dale, 1914-16) passes beneath this intersection and follows the trend of Route 9 for a few hundred feet.

Mileage (cont'd)

At the junction of Route 9 and Sleepy Hollow Road, turn left (west) on Route 9.

16.75 Turn north on the road to the Searsburg Reservoir and park out of traffic.

Stop 8. PRECAMBRIAN QUARTZITE AND LIME SILICATE GNEISS

On the northwest corner of this intersection is a small outcrop which together with the rock units at Stop 7 exemplifies several features typical of the boundary between the Cavendish Formation and the well-authenticated Precambrian rocks of the Mt. Holly Complex. This outcrop of blue-quartz quartzite of the Mt. Holly Complex has the attitude N.70°E., 90°. The presence of blue quartz is a characteristic feature of a number of the units of the Mt. Holly Complex.

A few hundred feet southwest of this intersection are outcrops of rusty weathering calc-silicate granulite beds. The east-northeasterly strike of these beds contrasts strongly with the attitude of the overlying Cavendish Formation (Readsboro and Heartwellville Schists of Skehan, 1961, pp. 45-63) exposed a few hundred feet to the east, whose attitude is N.15°E, 60°SE, and which were studied at Stop 7.

Two hundred feet downslope to the east of this blue quartzite outcrop may be seen the penstock aqueduct, the foundation of whose pedestals are on a well developed sequence of identical and related kinds of Precambrian rocks. Crawl under the penstock at one of the openings and proceed on foot in a northeasterly direction to the Deerfield River and rock-hop your way to the outcrops of dark biotite-muscovite-quartz schist cropping out on the east side of the river. These rocks grade up into biotite-albite-garnet-quartz schists which in turn pass upward within a short distance (Skehan, 1961, Pl. I) into the green (continuous with the beds of Stop 7) and black quartz-mica schist of the Heartwellville Schist.

The Searsburg Conglomerate is difficult to find at this locality but has been exposed in one outcrop south of Searsburg Reservoir and consists of elongate quartzite pebbles in a calcite-biotite-chlorite-quartz schist matrix.

Return to cars and proceed north to the Searsburg Dam for .7 mile on the unpaved road. Turn around in the field

Mileage (cont'd)

adjacent to the gatehouse at the dam. The Precambrian gneisses and schists exposed in the spillway of the dam are separated by only 300 feet from the Cavendish Schists in the Deerfield River below the spillway. Return to cars and proceed south to Route 9.

17.95 Junction of Route 9 and road to Searsburg Dam. Turn left (east) on Route 9.

18.05 Trace of the Precambrian-Cambrian boundary (noted above at Mile 16.55) is approximately at this location. Continue east on Route 9.

19.05 The high ridges to the north of the river and Route 9 are the green garnet schist of the Heartwellville units.

19.35 Bridge over Bond Brook of Stop 7.

20.25 Large outcrops of black quartz-mica schist of the Heartwellville Schist, on the left. The black and green beds of the Heartwellville Schist also outcrop from mile 20.35 to 20.90.

21.55 On the left may be seen the high cliffs of Stop 9.

21.65 Wilmington-Searsburg Township Line.

21.95 Medburyville Bridge. Make a U-turn and proceed west on Route 9 0.1 mile and bear right on an unpaved road. Proceed 0.35 mile to the old hotel beyond the Wilmington-Searsburg Township line and park off the road.

Stop 9. HOOSAC FORMATION, SEARSBURG CONGLOMERATE, READSBORO SCHIST, SHERMAN MARBLE AND HEARTWELLVILLE SCHIST.

Excellent exposure of cliffs of albite schist of the Hoosac Formation (formerly considered to be Readsboro Schist in Skehan, 1961, Pl. I) in contact with green and black schist unit of the Heartwellville Schist to the east. This albite schist is regarded as Hoosac Schist since it is now known to contain amphibolite similar to the Turkey Mountain Member. Traverse easterly across these beds to the contact with the coarse-grained albite schist of the Readsboro Schist enclosing layers of calcite marble of the Sherman Member. Proceed northeasterly to the outcrops of Searsburg Conglomerate exposed northeast of Medburyville and pictured in Skehan, (1961, Figs. 13 and 14, pp. 46-47 and described on pp. 45-49).

Mileage (cont'd)

Return to Route 9 and go west 4.20 miles to the junction of Sleepy Hollow Road.

26.15 Turn left on Sleepy Hollow Road. Proceed 2.5 miles to the junction of Sleepy Hollow Road with Route 8. Farrington Cemetery, the same as at Mileage 13.95, is on the left.

28.75 Turn left on Route 8 and proceed south 2.1 miles.

30.85 Stop 10. READSBORO SCHIST

Outcrops of dark muscovite-biotite-albite-garnet-quartz schist of the Readsboro Formation (Skehan, 1961, pp. 49-57) are exposed in the north fork of the west branch of the Deerfield River north of Heartwellville. These outcrops are immediately east of the inferred location of the Hoosac Thrust.

Proceed south on Route 8 a distance of 0.55 mile to the junction of Routes 100 and 8. Proceed easterly (left) on Route 100, 1.2 miles. Park off the highway near Lamb Brook 0.1 mile south of Stop 11.

32.60 Stop 11. HEARTWELLVILLE SCHIST

Walk back to the outcrop. Excellent road cuts in the dark schist of the Cavendish Formation (Heartwellville Schist, Skehan, 1961, Fig. 16, p. 60) at the type locality of the Heartwellville.

33.80 Retrace the route 1.2 miles to Routes 8 and 100. Proceed south on Route 8.

34.00 Heartwellville Center.

34.70 To the west of the highway in the grove of trees a quartz breccia is recognized and interpreted as fault breccia related to the Hoosac Thrust.

35.20 Dutch Hill Ski Area.

35.80 Heartwellville Lodge to the right.

36.60 The inferred location of the Hoosac Thrust between Heartwellville and Stop 12 lies west (to the right) of the highway. The ridge to the west is the Green Mountain core whose eastern part is flanked by the Cambrian Dalton

Mileage (cont'd)

Formation consisting of thin vitreous quartzite beds, schistose feldspathic quartzite and biotite-albite schist. The ridges to the east are comprised of the double decker overthrust sheets of the Cavendish units on the lower thrust and the Tyson-Hoosac units on the upper thrust.

39.60 Stop 12. HEARTWELLVILLE SCHIST, DALTON FORMATION AND STAMFORD GRANITE GNEISS.

Turn right (west) from Route 8 and go 0.7 mile to the home of Arthur Lincoln. Park off the road and in his yard and proceed up the hill to the large outcrops of garnet-chlorite-quartz schist of the Heartwellville Formation lithologically identical to the Pinney Hollow Formation (Tables 12 and 13, pp. 61 and 63). Traverse this section up slope, (down stratigraphically) to the contact of the Heartwellville with the Dalton Formation.

The Hoosac Thrust is interpreted as bringing the shale and graywacke facies of the Cavendish units to a position above the autochthonous rocks of the Cambrian beds which are traceable approximately three miles to the south to fossiliferous beds of the Olenellus zone of Clarksburg Mountain in North Adams discussed above. Proceed westerly to the contact of the Dalton beds with the Precambrian Stamford Granite Gneiss. Return to cars and return to Route 8. Turn right and proceed south toward Stamford on Routes 8 and 100.

41.25 Stop 13. VIEW AND PHOTO STOP

A view to the south along the Stamford Valley, underlain by Quaternary Alluvium which in turn may be underlain by Cheshire Quartzite as well as Cambro-Ordovician carbonate beds such as are exposed at Natural Bridge in North Adams. The steep western slope of Hoosac Mountain is developed above the easterly dipping Hoosac Thrust Fault, the trace of which is near the base of the slope. This slope may contain the traces of multiple thrusts which have been mapped by Norton in the Windsor quadrangle (oral communication, 1972).

Mt. Greylock, (el. 3,491 ft., the highest mountain in Massachusetts) comprised of marble interbedded in albite schist and green and dark muscovite-quartz-mica schist, looms up directly to the south. The Cheshire Quartzite and the Dalton Formation of the autochthonous sequence to

Mileage (cont'd)

the west of the viewer may be traced on the skyline in a southwesterly direction as they continue around the southerly plunging end of the Green Mountain Anticlinorium in the vicinity of North Adams and Williamstown. After this view, the field trip participants who are going south and east have several options. The junction of Route 2 and Route 8 is 5.35 miles to the south. The New York Thruway may be reached by following Route 2 west about 50 miles to the vicinity of Albany. The Massachusetts Turnpike may be reached by following Route 2 east to I-91 at Greenfield a distance of about 35 miles (driving time 50 minutes) and going south on I-91 to Springfield. Alternatively Route 2 may be followed west to Route 7 south which in turn meets the Massachusetts Turnpike at Stockbridge, Massachusetts, about 50 miles south of North Adams.

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Trip B-13

POLYMETAMORPHISM IN THE RICHMOND AREA, VERMONT

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SUMMARY

Rocks in the Richmond Area, Vermont consist mainly of wackes and phyllites with minor slates, quartzites, and amphibolites. These lithologies are divided into the Richmond Pond Phyllite and the Huckelberry Hill Wacke of the Pinnacle Formation, the Verdis Montis Amphibolite, and the Preston Pond Phyllite and the Duck Brook Wacke of the Underhill Formation. These units are correlated with the previously undivided Pinnacle and Underhill Formations in adjacent areas.

Graded bedding was used to indicate the way up in the section, which was preserved, along with evidence of six deformations. The sequence of deformations, as deduced by comparing the offsetting relationships of structures in single outcrops containing more than one structure, indicates that the area was folded, refolded, cleaved, the cleavage folded, kinked, and jointed, in order of decreasing relative age. The outcrop pattern is primarily second fold generation. The regional schistosity and the cleavage are the most commonly recognized structures. The folding of the cleavage and the kinking were minor events which were recorded only in the western part of the area, an area in which some of the joint planes are filled with basic igneous dikes. The first folding is believed to be Taconian, the second Acadian, and the kinking related to the Hinesburg thrust to the west of the area.

Recrystallization was associated with periods of folding and cleavage formation. The rocks were metamorphosed at the greenschist facies level each time, with the formation of biotite associated with the second period of folding being the highest level attained. A correlation of structure and metamorphism is combined to produce a tectonic sequence of deformational events for the Richmond area.

The purpose of this trip is to examine polymetamorphic assemblages in the Huckelberry Hill Wacke. Since many of the relationships between structure and metamorphism can be seen in hand specimens of this unit, it is suitable for field analysis. The wacke is dark green or drak gray in color depending upon whether pyrite + magnetite or magnetite along is present as an accessory phase. This difference governed the mineral assemblages associated with the fourth and final recrystallization. The three earlier recrystallizations, however, appear to have produced similar mineral assemblages throughout this unit.

environmental geology

Environmental Geology Cover page: Upper: Sanitary landfill, Randolph, Vermont. Lower: Resistivity study, Hinesburg delta, Hinesburg, Vermont. Photos by Arthur Huse, UVM Geology Department.

EG-1

MOUNT MANSFIELD TRAIL EROSION

Computerized statistical analysis of hiking trail erosion on a scenic area along the Green Mountains. Geology and vegetation of mountainous areas and their relationship to human recreational activity.

(The complete text of this paper will be available at the meeting in October.)

"Of the origin of Tah-wah-bebe-e Wadso--The Saddle Mountain--which became Lion Couchant, then Camel's Rump, later Camel's Hump; and its companion, Mount Mansfield, the Reverend Perrin B. Fiske speculated:

The Camel's Hump is there on high,
His head the sages think,
Is by the river's brink, where once
He ran to kneel and drink.
But stumbling in his thirsty haste
He threw his rider high,
And there lies Mansfield as he fell
A-staring at the sky."

From: Hill, Ralph Nading, 1949, The Winooski, Heartway of Vermont: Rinehart & Company, Inc., New York, p. 242.

Trip EG-2

FEASIBILITY AND DESIGN STUDIES: CHAMPLAIN VALLEY SANITARY LANDFILL

by

W. Philip Wagner and Steven L. Dean*

INTRODUCTION

In theory, solid waste disposal in Vermont has progressed from dumps to sanitary landfills, but in practice the differences between the two often are obscure. According to a recent review, "Over 90% of the small towns in Vermont dispose of their refuse in open dumps or substandard landfills"(Report of the Governor's Task Force, 1970). There is growing evidence that some of the better sanitary landfills are polluting (Thompson and Costello, 1972; Wagner et al., 1971; Wagner and Thompson, 1971). Although recycling eventually may solve the solid waste problems, sanitary landfilling is the only practical method presently available for Vermont.

This report is intended to illustrate that:

- knowledgeable landfill location and site evaluation can greatly reduce the chance of environmental degradation...

- sanitary landfills are not merely covered dumps, but in fact represent specially designed systems...

- short of recycling, there can be such a thing as a "good landfill", even in Vermont.

This is not a comprehensive account of all aspects of landfills. Emphasis is focused on pertinent, but commonly ignored geological and hydrogeological factors. The bibliography includes all publications reviewed in this project.

LANDFILL LOCATION

Much of the work presented here stemmed directly from a request from Paul Casey, Hinesburg Sand and Gravel Co., Inc., for help in designing a landfill that absolutely would not degrade the environment. Thus, the problem began, at least in a general way, with a given location near Burlington. For a private operator, a public official, or a planner faced with the initial problem of locating a suitable landfill site, the procedure to be followed would be much the same as used here. The Appendix includes a check list for evaluation of different sites. The following discussion deals with environmental guidelines for landfills.

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Sanitary landfills can be located in almost any place, but if financial costs for protecting the environment are to be minimized, it is desirable to recognize and take advantage of certain natural characteristics of the land. The problem, simply stated, is to identify criteria for locating landfills in Vermont. If meaningful, such criteria will aid rather than hinder landfill development. Sound guidelines for locating landfills will make good economic as well as environmental sense.

The logical way to develop criteria is to consider previous studies on the subject. Literature dealing with sanitary landfills is extensive. In some places certain criteria have been developed, but most publications relate studies of individual landfills. Some aspects of studies elsewhere may not be directly applicable to Vermont due to differences in topography, climate, soils, or rocks. On the other hand, similarities in reports from diverse places indicate that there are some universal "truths" that cut across political boundaries. By combining information from various studies it is possible to develop criteria for locating landfills according to substrate and cover materials. Depending on whether the substrate is relatively permeable or impermeable, the following criteria can be identified:

1. Permeable substrate, generally sand and gravel, with:
   a) minimum 1000 feet to nearest perennial stream
   b) minimum 30 feet of <u>dry</u> substrate below landfill base
   c) maximum 10% slope

2. Impermeable substrate, generally certain glacial tills and some lake or marine bottom sediments, with:
   a) minimum 200 feet to nearest perennial stream
   b) minimum thickness of 5 feet of substrate below landfill base
   c) maximum 10% slope
   d) minimum 6 feet of <u>dry</u>, permeable material overlying impermeable substrate
   e) leachate control and treatment

The current trend nationally is toward sites with impermeable substrata. In such sites leachate is either prevented from leaving the landfill, or moves at such low velocities that it undergoes optimum purification by chemical and biochemical reactions, filtering, and dilution. Landfills with permeable substrate may be suitable for certain kinds of waste material not likely to cause environmental degradation.

As for cover materials, both impermeable and permeable soil covers have been used elsewhere. The former has the advantage of repelling surface water, thereby minimizing leachate generation, but retarding gas release. The latter promotes upward escape of gas but also allows for surface water infiltration leading to increased leachate production. A formula of 80% well-graded gravel,

10-15% sand, and 5-10% fines provides a relatively impermeable cover that, with specially designed gas vents, offers optimum conditions for controlling leachate production, gas diffusion, rodents, flies, and frost heaving. In addition, such material can be compacted and can support heavy vehicle traffic. Thus, site location considerations should include, in addition to substrata conditions, the availability of sufficient volumes of cover materials which will offer the benefits outlined above. In Vermont, the natural deposits most closely resembling the ideal cover material are certain glacial tills and glacial gravels. In most cases, however, cover material probably will have to be specially prepared by mixing materials of different grain size.

SITE EVALUATION

Location, Topography, and Drainage: The proposed site in question involves about 25 acres of relatively impermeable soils, approximately 3 1/2 miles southeast of Hinesburg Village, in the Town of Hinesburg (Figure 1). The site is situated in the foothills of the Green Mountains in an area of gently rolling topography. Elevations of the land surface at the vicinity of the site range from below about 500 feet to about 420 feet over long, gentle slopes (Figure 2).

Drainage in the area is westerly as part of the Lewis Creek drainage basin. Hollow Brook, the perennial waterway closest to the site, is almost 2000 feet to the north. A small intermittent stream is located along the south and west margins of the landfill area. Although the surface waters in Lewis Creek are intended to be classified as "B" (suitable for drinking with treatment), samples taken in 1956 indicated class "C" (unsuitable for drinking) coliform levels (Vermont Department of Water Resources, 1968, p. 16).

Elevations at the landfill site are above flood levels from any streams. However, Hollow Brook to the north is actually at a higher level than the site. Surface flooding of the site from Hollow Brook is prevented by extensive, high deposits of gravel between the landfill site and Hollow Brook. These deposits should be partially preserved from commercial gravel excavations to prevent southward diversion of Hollow Brook through the landfill site.

Soils: From the point of view of soils and topography throughout Chittenden County, the South Hinesburg area is considered as having good potential for sanitary landfills (Sargent and Watson, 1970). However, the detailed soils map of the area by the Soil Conservation Service (Figure 3) shows some limitations for landfills. A summary of the pertinent aspects is presented in Table 1.

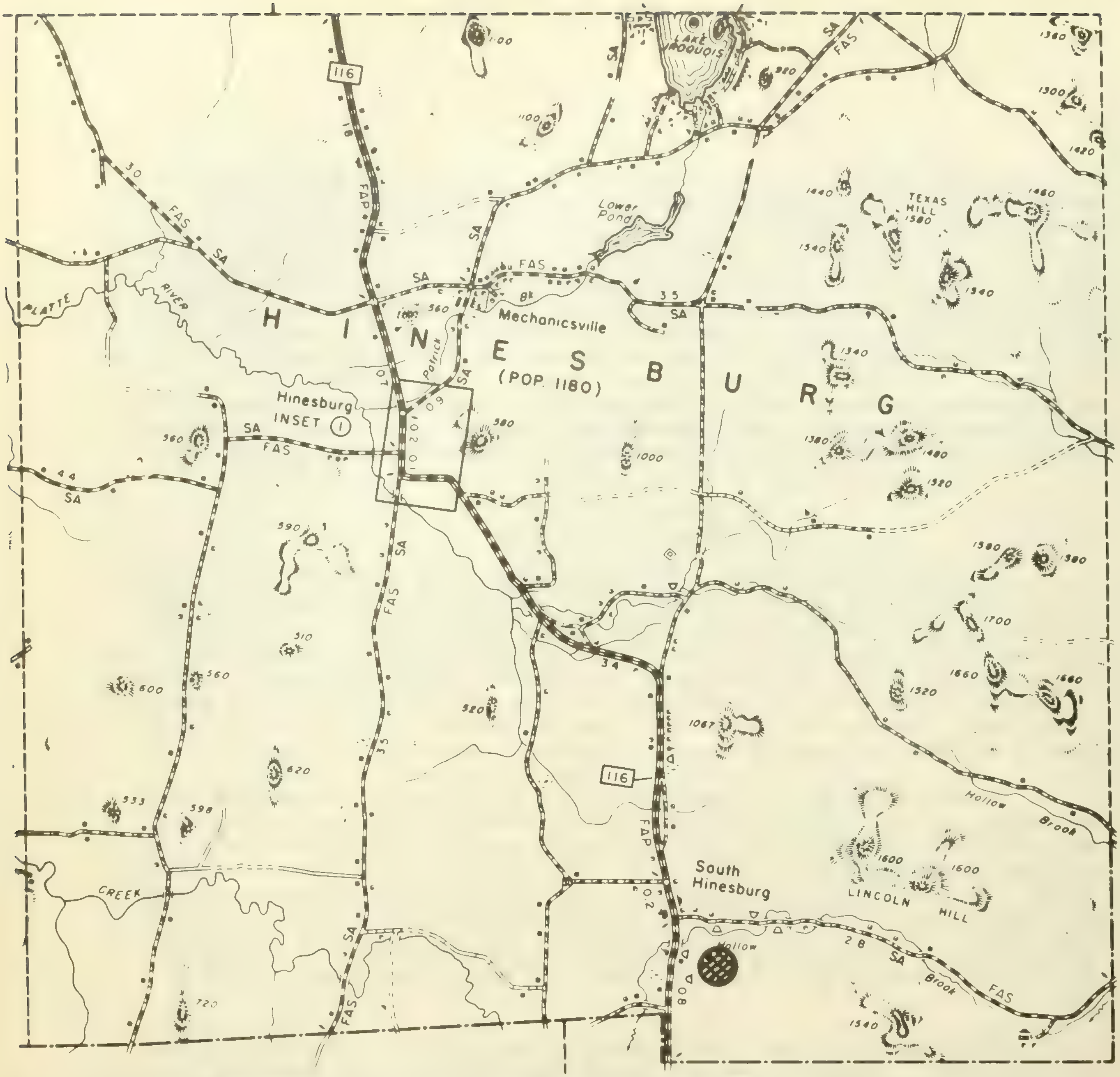

Figure 1: (top) Location of site on County Highway Map(diagonally-ruled circle.

Figure 2: (bottom) Topography at site and vicinity (diagonally-ruled circle).

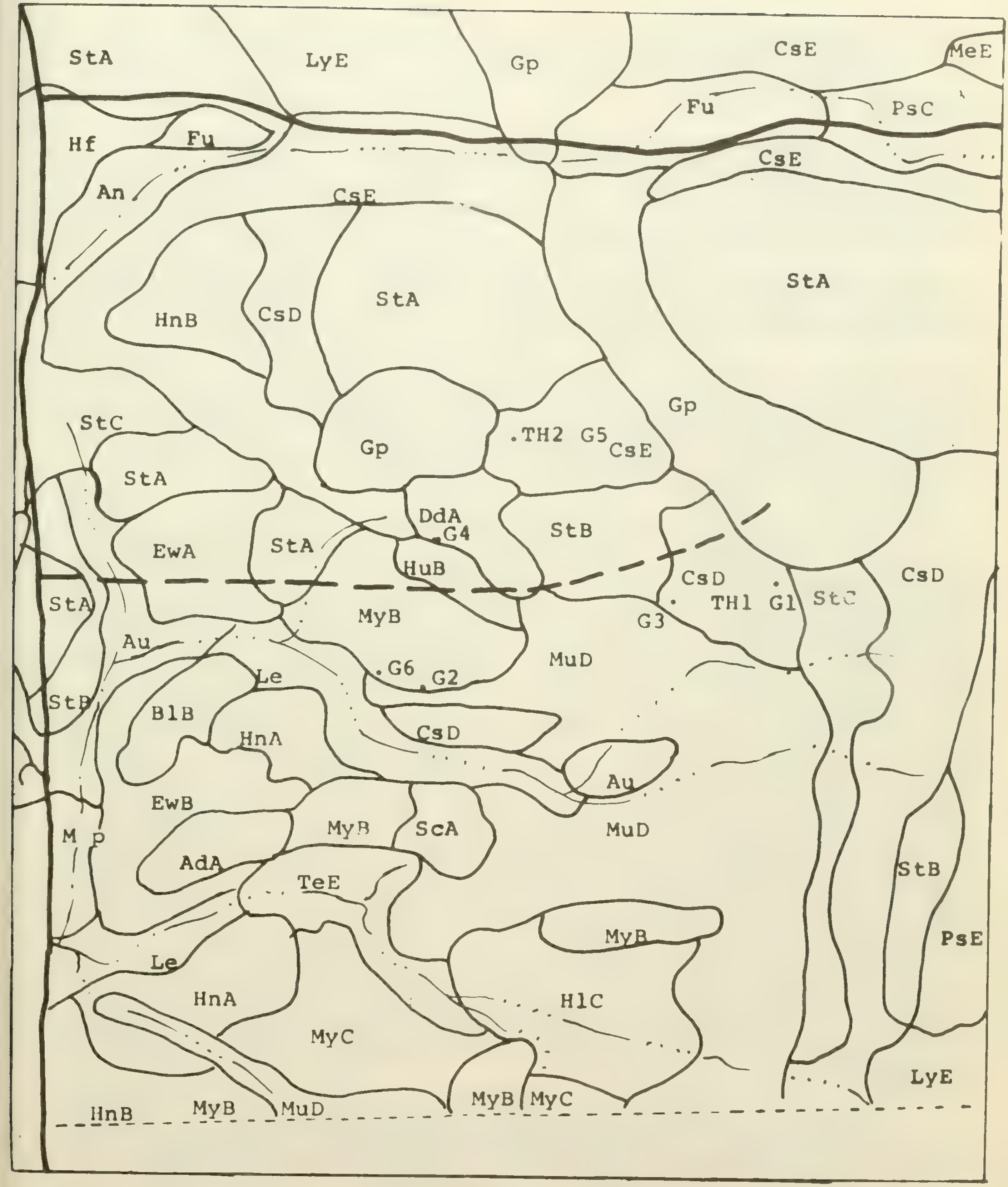

Figure 3: Detailed soils map of the landfill area by Soil Conservation Service. Units are explained in Table 1. TH = test hole; G = geophysical test.

Table 1: Soils at the landfill site and vicinity.

Soil Type (map symbol)	Slope (%)	Limitations
AuGres fine sandy loam (Au)	---	high water table
Colton and Stetson soils (CsD)	20-30	steep slopes
Duane and Deerfield (DdA)	0-5	high water table
Enosburg and Whately (EwA)	0-3	high water table
Hinesburg fine sandy loam (HnB)	3-8	low permeability and strength
Munson and Belgrade silt loams (Mud)	12-15	high water table
Munson and Raynham silt loams (MgB)	2-6	high water table
Stetson gravelly fine sandy loam (StB)	5-12	steep slope

In the immediate area of the landfill the dominant soil types have problems with seasonal high water tables due to low permeability. It should be pointed out that such water tables are "perched" types due to the retention of precipitation at and near the surface. This problem, unlike deeper ground water, can be overcome easily by appropriately designed drainage controls.

The amount of water that collects on the land surface at the site can be estimated.[1] Due to the highly permeable character, the irregular topography, and low ground water table of gravel areas adjacent to and uphill from the site (north and east), surface waters readily infiltrate the gravelly soils or are naturally diverted around the site. Thus, the water that collects on the impermeable surface at the site is derived primarily from rain and snow directly on the site itself. Of the 30-40 inches of annual precipitation in the area, about half is lost by evapotranspiration. The remaining 15-20 inches, representing 31-42 acre-feet over the 25 acres of the site, constitutes surface runoff. Due to the seasonality of precipitation and evapotranspiration, larger amounts of water are expectable during the spring and fall than other periods. The amount of water at the site due to snow melt is about 10 inches (water equivalent), or nearly 21 acre-feet, very little of which is lost by evapotranspiration. The non-snow precipitation of 20-30 inches, on the other hand, is reduced by about 90% by seasonally high evapotranspiration to about 2-3 inches or about 4-6 acre-feet. The problem of poor surface drainage is at least three orders of magnitude greater in the spring than in the rest of the year. This can be reduced by snow removal to negligible amounts. During the remainder of the year slightly less than 2,000,000 gallons of water will enter the site. Initially, most of this water will be diverted westward, away from the landfill operation.

Geology: Bedrock in the area is completely buried by unconsolidated materials. Regional geologic studies, however, indicate that the buried bedrock consists of the Underhill formation, a micaceous schist. The schist is impervious to water except where

1 Robert Hendricks, U.S.D.A., provided meteorological data and helped with estimations.

joints (cracks) have developed. In this region joints are less abundant than elsewhere. As a result, ground water movement in bedrock is highly restricted and, therefore, less sensitive to pollution than usual.

The deposits overlying bedrock largely determine the environmental suitability of the landfill. General geologic information shows the landfill site is in an area of former lake bottom where fine-grained sediment was deposited. The gravel deposits immediately north of the site are in a deltaic deposit formed in the same lake. East of the site at the surface, and buried beneath much of the fine-grained sediment at the site itself, are gravel deposits produced by the ice sheet in the area. Over much of the area glacial till is expectable beneath the gravels and fine-grained sediments, and directly over bedrock.

Detailed information on subsurface geologic conditions has been obtained by drilling and by geophysical (seismic and resistivity[2]) testing. Information from the tests, which are located on Figure 3, is presented in cross-sections in Figure 4. Bedrock ranges from 50 to 100 feet beneath the land surface, with greater depths in the deltaic deposits north of the landfill site. The slope of the bedrock has a distinct westerly and southwesterly component, somewhat similar to the present land surface. Buried till is present in the eastern part of the area (profile A, Figure 4) at sites TH1-G1 and G3, but is not evident at other sites. A thick gravel layer is the dominant feature of the subsurface materials. This gravel is overlain in most places at the site by up to 20-30 feet of the fine-grained lake sediments.

Ground Water: As previously mentioned, perched water collects at and near the surface of the lake sediments at the site. Whether or not this constitutes ground water is a semantic and academic question. Such near-surface waters are not generally used for water supplies. As pointed out previously, this water can readily be controlled. Water at greater depths in the ground, on the other hand, constitutes a natural resource that must not be contaminated by the landfill. Testing has shown that the gravel deposit buried beneath the fine-grained surface sediments contains ground water and, therefore, constitutes an aquifer. Water table slopes (Figure 4) indicate that recharge to this aquifer is provided by Hollow Brook (an influent stream) and undoubtedly to a lesser extent by percolating surface waters in the gravel deposits north and east of the site. Ground water movement is westerly to southwesterly. Changes of the level of the water table are expectable with time, primarily at different seasons of the year. Measurements of the water table depth in the test holes show only slight changes to date. Based on statistical analyses[3] of four gravel wells monitored by the Vermont Department of Water Resources, we have projected probable future changes in the water table in the test holes. These projections show that ground water remains well below the surface at all times, with seasonal fluctuations no more

[2] Resistivity data provided by Arthur Huse.

[3] Statistical work by Steven Pendo.

than about 10 feet.

From the viewpoint of ground water contamination it is important to note that the landfill site is not a recharge area for the gravel aquifer. Significant downward movement of leachate through the fine-grained sediment is not expectable. Percolation tests in such materials have shown exceedingly low rates of movement (Mullen, 1972; Waite, 1971). Thus, with special precautions to control and monitor leachate movement, ground water contamination can be prevented.

Miscellaneous: A variety of aspects deserve brief mention.

1. Biota: The immediate area of most of the landfill site has been actively farmed until the present time, so that no natural plant species are endangered. Along the periphery of the landfill on all but the north and northwest sides are common species of mixed hardwood and softwood trees, grasses and sedges. Animals in the area are likewise common species. No damage to ecologically fragile or otherwise unique biota is likely to occur.

2. Forest reserves: The landfill site mostly lacks timber except along the eastern fringe. The site is on the margin of the productive forest area of the Green Mountains, with soils rated fair at best for potential forest productivity (Gilbert, 1970).

3. Agricultural reserves: According to Carlson et al. (1970, p. 3), the landfill area is in a classification noted as "..the least suitable of all land now being used for agriculture in the county." Moreover, the area's present agricultural land use is considered by the same authors to be marginal to poor.

4. Natural areas: The site has no known value as a natural area deserving protection for biologic, geologic, archaeological, or other natural characteristics.

5. Aesthetics: View of the landfill site is blocked by high banks of gravel to the north, by the Green Mountains and tree cover to the east, and by a fringe of trees along the south and southwest margins. The only open view of the site is from the northwest and west. This will be remedied by tree plantings. Thus, complete privacy for the operation will be provided from all public vantage points.

6. Erosion: Erosion is not a problem in the area of fine-grained soils due to the soil cohesion and particle size. In gravel soil areas, only artificial slopes greater than about 65% show evidence of instability and erosion.

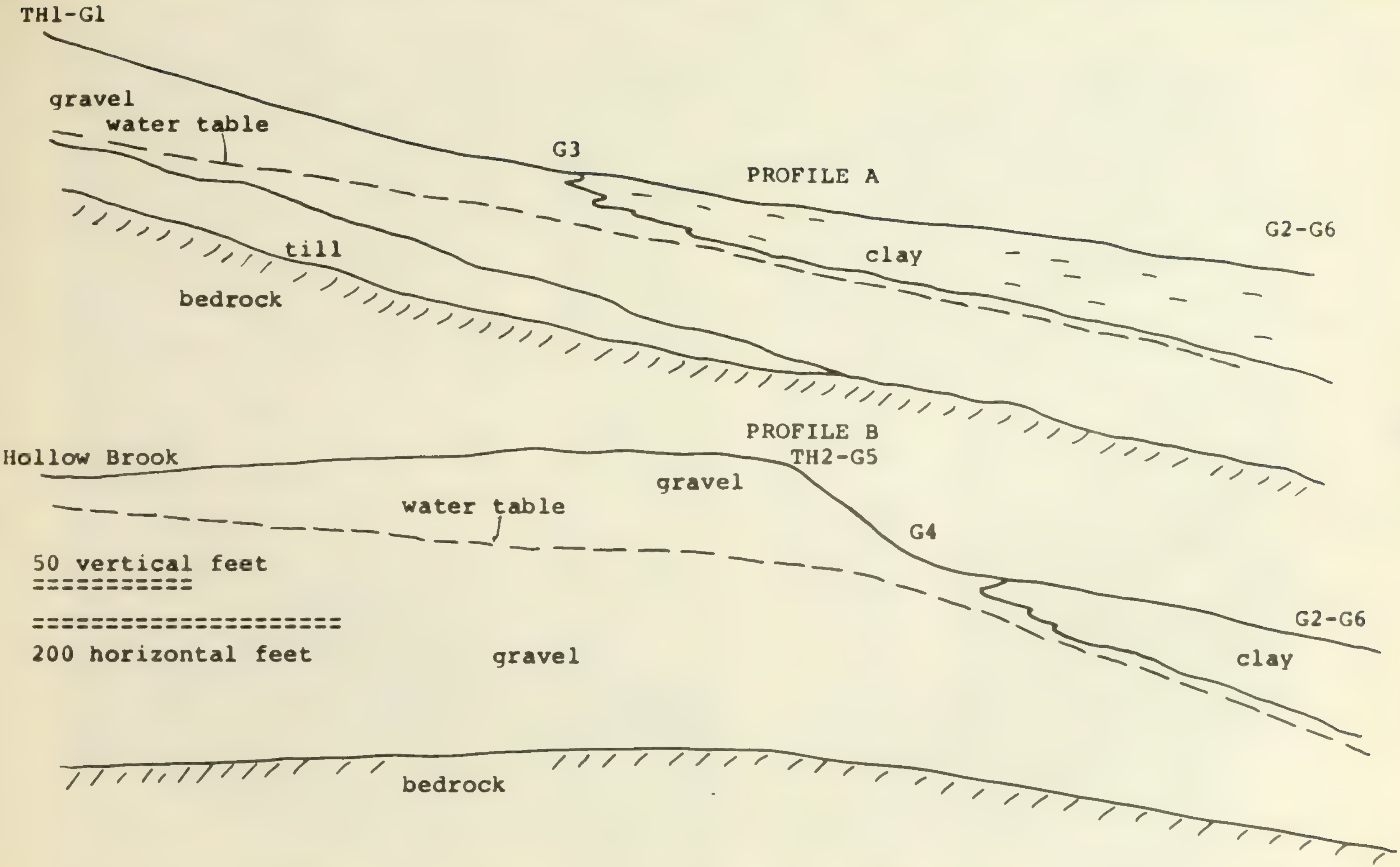

Figure 4: Approximate east-west (upper) and north-south (lower) cross-sections of landfill area. See Figures 3 and 5 for locations of tests.

DESIGN AND OPERATION

The sanitary landfill here proposed involves a combination of trench and area methods, utilizing impermeable base and cover materials, and artificial leachate and gas movement controls. Diagrammatic aerial and cross-section views of the landfill are given in Figure 5. Initially, non-bulky refuse will be placed in a trench system oriented north-south. After the trenching operation is completed, a superposed area-fill type of landfilling will commence. Based on an average total fill thickness of 50 feet with a waste to cover ratio of 4:1 and a 1000 lb/yd^3 density for compacted fill, the anticipated life span of the operation is about 22 years per 40,000 persons served. Bulky, non-putrescible items will be handled separately in the areas shown in Figure 6.

Cover material for the operation will be an artificially pre-mixed formulation of 80% well-graded gravel, 10-15% sand and 5-10% fines. Sand and gravel for the cover will be taken from the nearby commercial operations. Fines for the cover material will be obtained from the silt-clay layer at the site itself. Sufficient volumes of cover material are available for at least 100 years operation per 40,000 persons.

Effluent Control: Due to the impermeable nature of the cover, little or no leachate is expected from the landfill. However, special design conditions are recommended to insure that no ground or surface water pollution can be caused by leachate. Fill-trench floors in the fine-grained sediment will be sloped and veneered with gravel to direct drainage from the fill-trench system to a filter-storage trench on the northern margins of the fill. Berms will divert surface waters away from the site and away from the filter-storage trench.

A pump system will draw leachate through an underdrain in the filter-storage trench and transfer the leachate to steel storage tanks located at the western end of the site. The landfill operation will begin at the eastern margin of the landfill-trench system. At first only a small portion of the total site will be developed, the actual size depending on the size of the population served. Assuming wastes are collected for 40,000 persons, the trenching required will involve about 3 acres per year. The volume of leachate, based on infiltration of rain and snow removal, should be less than 250,000 gallons the first year and 500,000 gallons the second year. The steel tanks will hold an aggregate volume of 30,000 gallons, which when combined with the filter-storage trench capacity of about 500,000 gallons, will provide storage in excess of the amount expected for the first and second years of operation. At the end of that time sufficient data will be available to plan for increased storage capacities as necessary.

Depending on the chemical quality of the leachate collected, it may be pumped from the steel tanks to the distribution line of

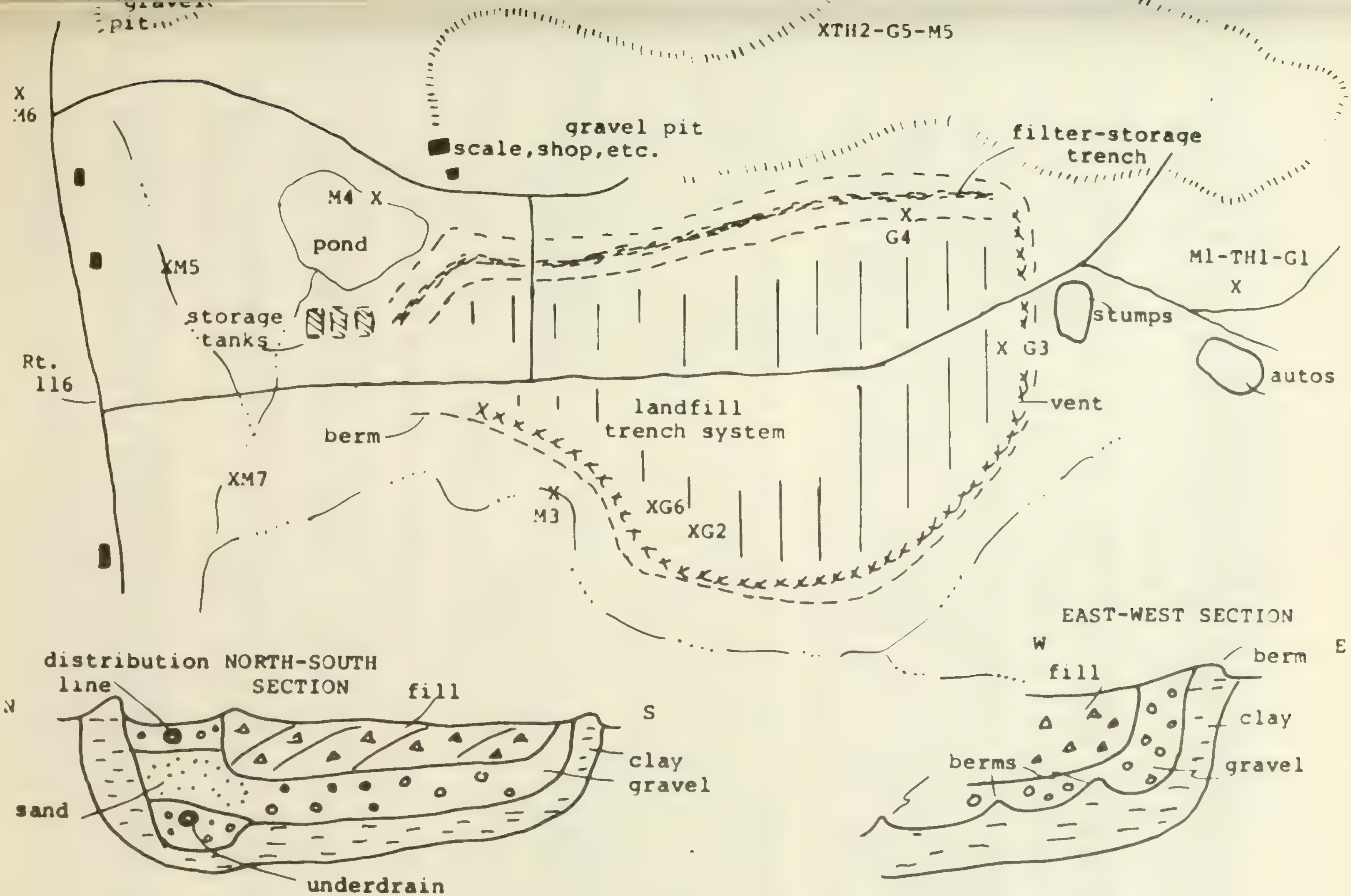

Figure 5: Schematic plan and section views of landfill area. M = monitoring stations; G = geophysical tests; Th = test hole.

the filter-storage trench for filtration (Figure 5). Alternatively, the leachate may be chemically treated. Release of treated leachate will be effected by pumping it to the gravel area northeast of the landfill. Here the large thickness of dry gravel will provide further filtering.

Gas Control: Gases produced in the landfill will be transmitted through the gravel on trench floors in an up-slope direction toward the eastern and southern margins. There the gases will be released to the atmosphere through the gravel vent.

Monitoring: Although elaborate steps will be taken to guard against water pollution, monitoring stations are to be used for periodic sampling of natural surface and ground waters at sites shown on Figure 5. Ground water will be monitored by sampling from perforated pipes installed in the test holes. Periodic checks of the ground water table elevation will be continued. Finally, close supervision will be made of the leachate quantity and quality in the steel tanks and in piezometers installed in and below the filter-storage trench.

Analysis of the biochemical quality of ground and surface waters will be guided by the quality of the leachate. Samples from all check points will be taken at least three times per year and at more frequent intervals from leachate storage facilities as required.

Miscellaneous:

1. Litter control: Snow-fencing erected on periphery of trench in operation.
2. Vandalism control: Two full-time attendants during operation; cyclone fence along periphery of landfill with locked gates during non-operation hours.
3. Fire control: In addition to benefit of cover material, pond adjacent to landfill can be used for water supply for fire fighting.
4. Access roads: All-weather, 24 foot wide, asphalt surfacing with grades less than 7%.
5. Buildings: Existing weigh scale station, and maintenance and vehicle storage sheds will be utilized (Figure 5).
6. Personnel facilities: Toilet and water supply facilities available in scale house.
7. Clearing and grubbing: Not necessary.
8. Rules and regulations will be posted as follows:
   a. No private use.
   b. All operations supervised during specially designated times.
   c. No salvaging without permission of owner.
9. Method of handling and compacting waste: Refuse will be dumped at toe of working face and spread to a 1000 lb. density with continuous spreading and compacting.
10. Site reclamation: Soil cover material at site will be stockpiled along margins of site for resodding upon completion of landfill.

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APPENDIX: Site Evaluation Considerations for Landfill Location

Economic Factors:

Initial-

acreage
estimated cost per acre
estimated access road cost
estimated site clearing cost
estimated site modification cost
estimated building cost
estimated engineering costs
estimated equipment costs
estimated fencing costs

Annual-

salaries and benefits
equipment operation
maintenance and repair
snow removal
depreciation
amortization of initial costs
administrative overhead
cost per capita

Other-

reclamation
recycling-distance from population centroid

Social Factors:

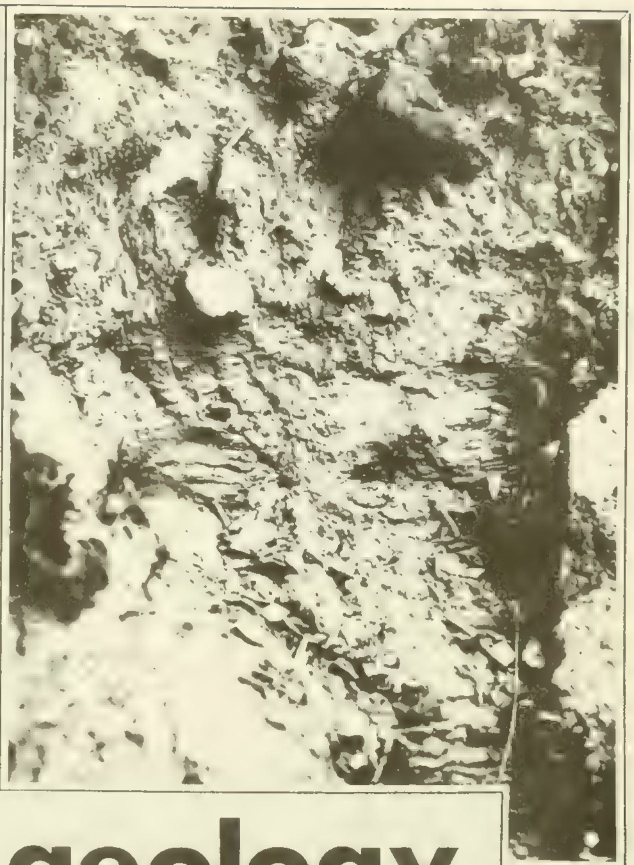
prevailing winds (incineration; dust; odors; noise)
aesthetics
present landuse on site
present landuse adjacent to site
landuse plans and zoning
fire protection
traffic flow congestion and safety
road conditions leading to site

Environmental:

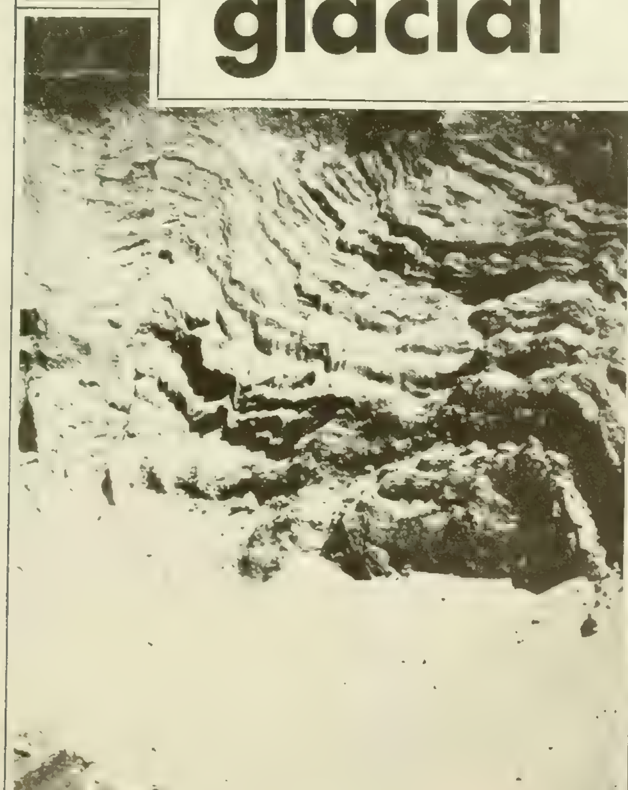
site volume
site longevity
substrate character and thickness
cover material character and volume
bulky item space
distances to perennial streams, and floodplains
slope
groundwater depth and flow direction

Environmental: (continued)

gas control
surface water control
distance to nearby wells
monitoring
near present or future sewage treatment plant



glacial geology



"Of all the phenomena of drift none have been more difficult to explain by any theories in vogue among geologists, than these trains of angular bowlders. To make water the sole agent, as some theories do, is the most unsatisfactory; for this could not alone have torn the blocks from their parent bed, and if it had been able to carry them forward at all, it must have rounded them. The most plausible resort would be to glaciers; but the nature of the surface over which the trains have been strewed, forbids the idea of a glacier. Common icebergs are no more satisfactory; but if we suppose islands capped with ice, and this occasionally torn up by the waves, and carried forward with fragments of rock in their under side, torn off from the islands and dropped along the way, or perhaps ice-floes in like manner frozen to the shore and torn off and urged along the coast, there is some plausibility in the explanation."

Edward Hitchcock, 1861
Geology of Vermont, v. 1, p. 65.

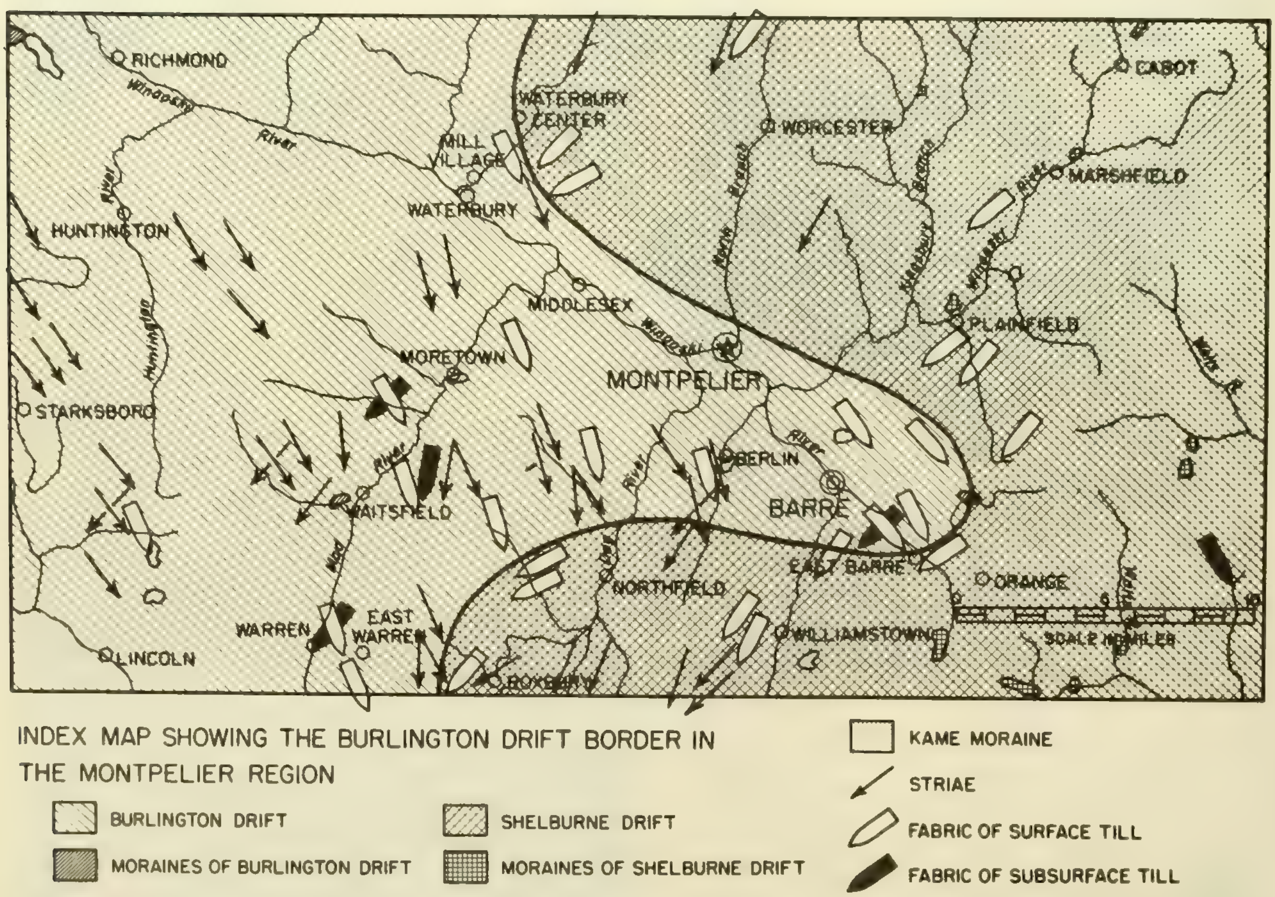


Figure 1. Map showing the Burlington drift border in central Vermont (from Stewart and MacClintock, 1969, fig. 15, published by permission of Dr. Charles G. Doll, Vermont State Geologist)

Trip G-1

GLACIAL HISTORY OF CENTRAL VERMONT

by

Frederick D. Larsen, Department of Geology
Norwich University

Introduction

The area traversed on this field trip lies on the Barre, East Barre, and Montpelier 15' U.S.G.S. topographic maps in central Vermont. The terrain is underlain by eugeosynclinal rocks which range in age from Ordovician to Devonian. The rocks were tightly folded and intruded by granite during the Acadian orogeny 380 million years ago (Naylor, 1971). Erosion has produced, over much of the area, a crude trellis drainage pattern which is characterized by alternating linear ridges and subsequent valleys which trend north-northeast. Drainage passes via the Stevens Branch and the Dog River northward into the Winooski River, a major superposed stream, which flows west-northwest through the Green Mountains to Lake Champlain.

During the Pleistocene central Vermont was probably completely covered several times by continental ice sheets, however, there is no clear evidence which supports multiple glaciation as it is known in the Midwest. The last ice sheet reached a maximum extent about 19,000 to 20,000 years ago on Martha's Vineyard (Kaye, 1964). Near Middletown, Conn., a readvance of the ice occurred before 13,000 years ago (Flint, 1956), and the Highland Front moraine was constructed in southern Quebec about 12,700 years ago (Gadd, 1964). These facts have led Schafer (1967) and others to conclude that retreat of the active ice margin in northern New England was very rapid (1000 ft/yr) and that removal of the ice took place by regional stagnation or downwasting. Lack of moraines and ice-shove features in central Vermont implies that downwasting was the dominant process during deglaciation.

Recently, the work of Stewart (1961), and Stewart and MacClintock (1964, 1969) has resulted in the controversial identification of three drift sheets in Vermont. From oldest to youngest they are: (1) Bennington drift, (2) Shelburne drift, and (3) Burlington drift. Separation of the drift sheets was made on the basis of striations and till-fabric studies which indicate that the Bennington and Burlington drift sheets were formed by ice moving from the northwest, whereas the Shelburne drift was oriented to the northeast. The relationship between the Burlington and Shelburne drift sheets in central Vermont as visualized by Stewart and MacClintock (1969) is shown in figure 1. One of the purposes of

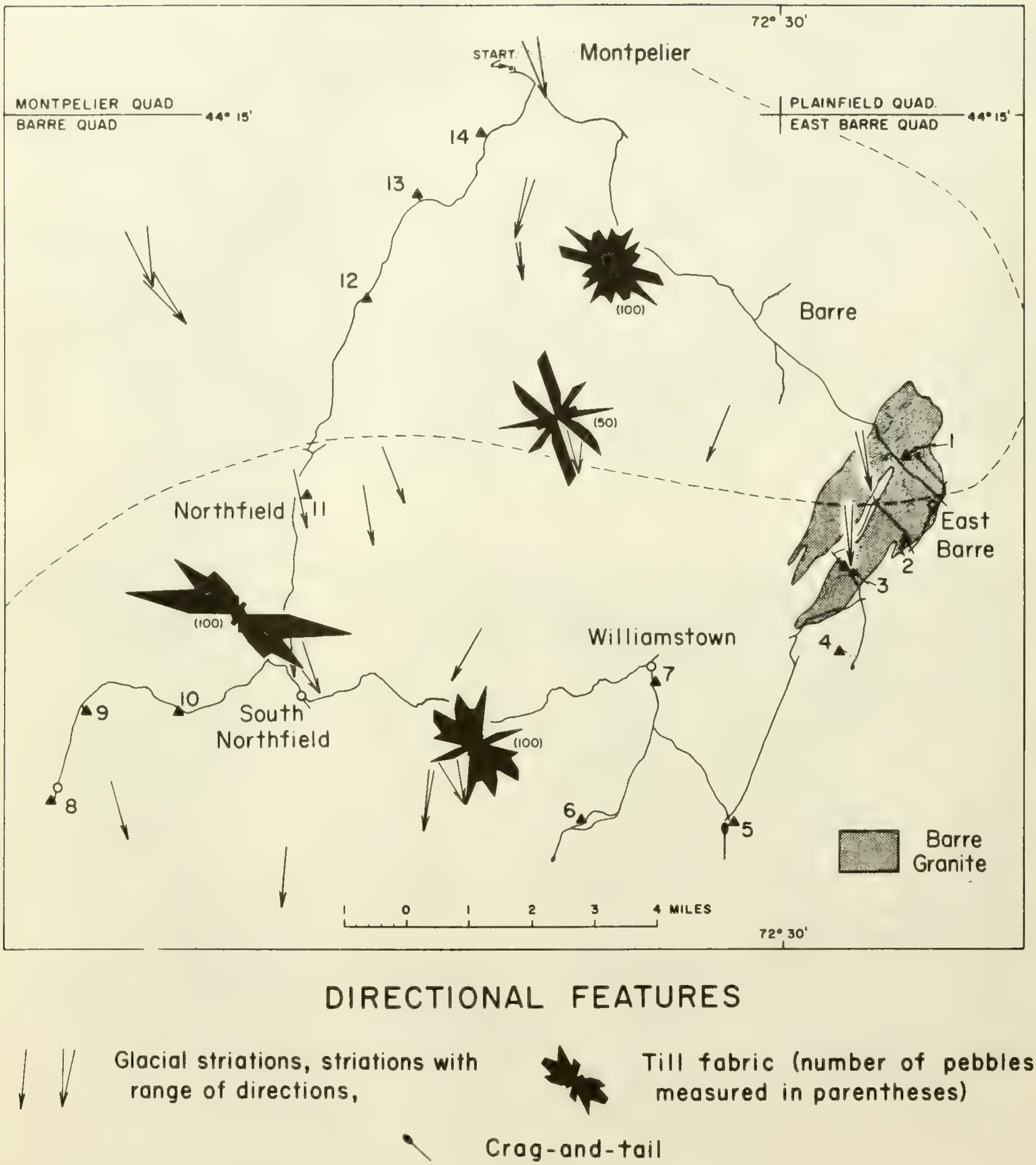


Figure 2. Glacial striations, till fabrics, and crag-and-tail feature in study area. Solid triangles represent field trip stops. Dashed line represents border of Burlington drift (compare with fig. 1). Directional features measured by F.D. Larsen, J.M. Ayres, D.A. Howard, J.G. Kvelums, S.A. Lawler, D.W. MacCormack, R.P. Magnifico, V.R. Sosnowski, and Squier.

this trip is to inspect, in the field, the validity of the relationship between the Burlington and Shelburne drift sheets.

Acknowledgements

This work, which is still in the reconnaissance stage, was originally developed as a doctoral problem but lack of funds prevented its pursuit for that purpose. The original impetus for this study arose from Douglas Selden, Norwich Univ., '65, who wrote a paper on the history of the Dog River valley. Selden discovered four major terrace levels that can be related to a sequence of proglacial lakes that formed during deglaciation. Several students at Norwich Univ. have contributed to this study through projects in glacial geology. Eugene Rhodes, Univ. of Massachusetts, donated his services as a field assistant for three weeks during the summer of 1967. Dr. Joseph H. Hartshorn, Univ. of Massachusetts, and Dr. Barrie C. McDonald, Geological Survey of Canada, have contributed ideas to this study.

Advance of Ice

During the last major advance of Wisconsin ice in central Vermont, movement was to the south and southeast across rugged terrain with relief on the order of 1000 to 2000 feet. In the area shown as Shelburne drift (fig. 1), mapping of striations, till fabrics, crag-and-tail features, and an indicator fan derived from the Barre pluton, suggests ice movement to the south and southeast and not to the southwest as postulated by Stewart and MacClintock (1964, 1969). Striations and till fabrics mapped by glacial geology students at Norwich Univ. are shown in figure 2 (compare with fig. 1).

Indicator Fan: An indicator fan based on pebbles derived from the Barre pluton was mapped during the summer of 1967. The first 100 pebbles encountered at each of 57 till localities were collected, washed, and, if necessary for identification, cracked open. The bulk of the pebbles were of metamorphic provenance comprising slates, phyllites, quartzites, and schists from the Waits River, Gile Mountain, Missisquoi, Stowe and Northfield Formations. However, 3 to 55 percent of the pebbles were of Barre-type granite, that is, light to medium gray granite with fine to medium texture. The percentage of granitic pebbles was plotted on a map and contoured with "isopers" (lines of equal percent) (fig. 3). The apparent long axis of the indicator fan trends toward S 15° E. Granitic pebbles which lie north and west of the 10 percent isoper represent a background count, and are assumed to have been derived from granitic bodies at Adamant, Woodbury, Hardwick, and unknown localities. Granitic pebbles derived from the Knox Mountain pluton, located to the east and northeast of the Barre pluton, are undoubtedly mixed with those from the Barre pluton. Since the color and the tex-

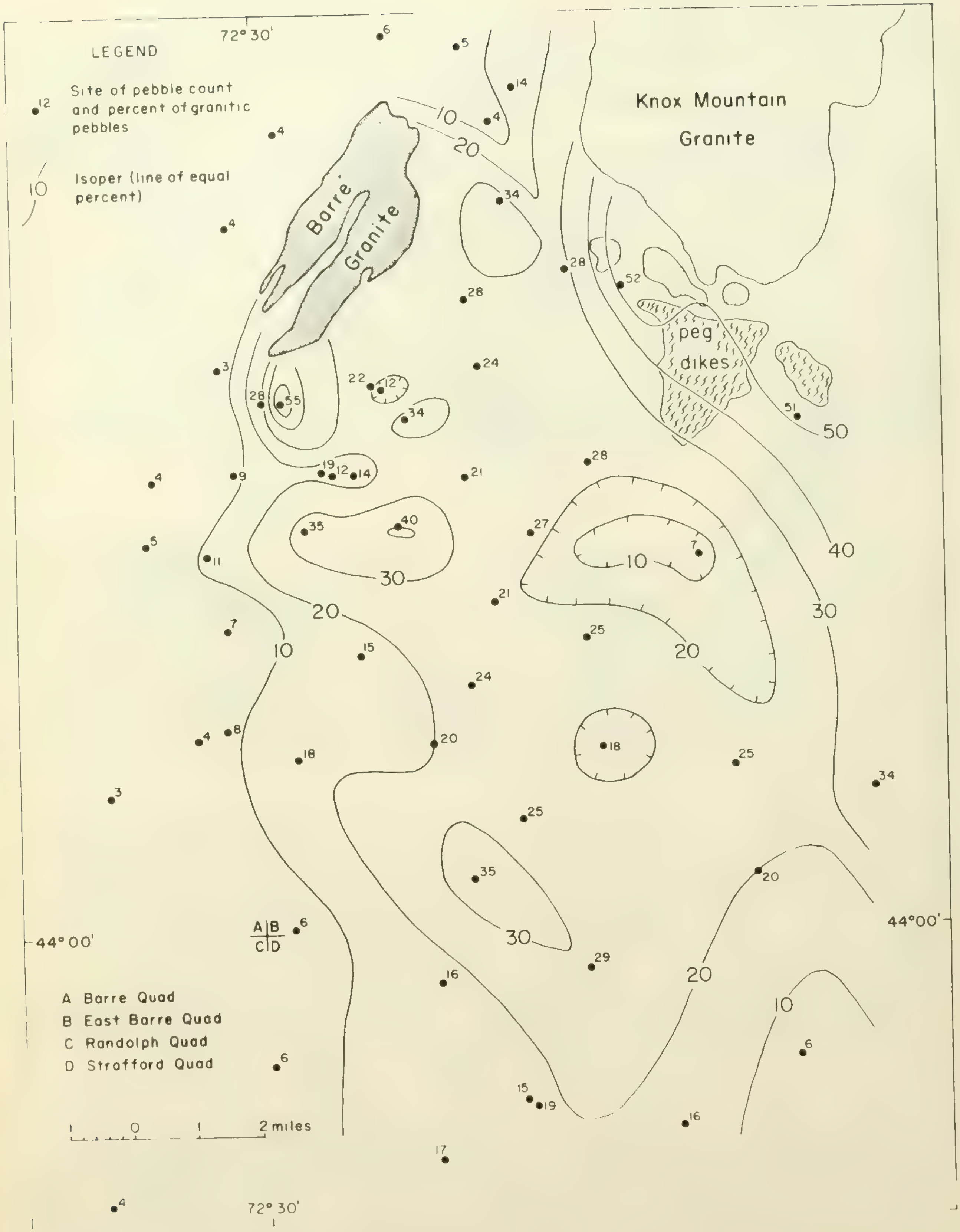


Figure 3. Indicator fan of pebbles from the Barre Granite (location of bedrock exposures from Murthy, 1957).

ture of granite from the two plutons is similar in appearance it is not possible to readily distinguish the source of pebble-size clasts. The Barre Granite is relatively homogeneous in texture and has few distinguishing features. In contrast, the Knox Mountain Granite is cut by numerous pegmatite dikes, and often contains garnets of pinhead size. Therefore, it is possible to identify the source of some of the larger erratics on the basis of features other than color and texture.

Boulder Train: East-west traverses in the area south of the Barre pluton indicate that there is a sharp line separating terrain with few granitic erratics on the west from terrain with numerous granitic erratics on the east. This line trends due south from the westernmost bedrock exposures of Barre Granite and roughly parallels the 10 percent isoper on the indicator fan. Although detailed mapping of granitic erratics is incomplete, the concentration of erratics is high in a north-south zone 0.5 to 2 miles wide and 10 miles long, and appears to decrease eastward over the next 3 to 4 miles, at which point erratics derived from the Knox Mountain pluton increase in numbers. A line representing the westernmost occurrence of granitic erratics with pegmatite dikes and/or pinhead garnets extends S 5° E from the westernmost exposure of Knox Mountain Granite.

It appears that, extending due south from the Barre pluton, there is a boulder train within a larger indicator fan which is defined by pebble counts, and which trends S 15° E. If this is true, I suggest that the first glacial erosion of the Barre pluton was by an ice sheet moving to the southeast. At this time, only pebbles and small erratics were being eroded. At a later time, when erosion had cut deeper into the pluton to pluck out large erratics, movement of the ice was due south. This suggestion of shift of movement from southeast to south has a precedent in diagrams of other Vermont indicator fans. As shown by Flint (1971, p. 178), indicator fans of Craftsbury Granite and a quartzite at Burlington have a long boundary stretching southward from a source area and a short boundary on the southeast side. This pattern may best be explained by a gradual shift in direction of movement from southeastward, as the ice sheet built up, to southward when the ice sheet reached a maximum thickness. Whatever the cause of this apparent or real discrepancy between the axes of the indicator fan and the boulder train, there is no evidence of major ice movement to the southwest in the vicinity of the Barre pluton as suggested by Stewart and MacClintock (1969).

Deglaciation

Downwasting of ice in central Vermont first witnessed the emergence of the Green Mountains as linear rows of nunataks. Evidence of vigorous fluvial erosion during the early stages of deglaciation comes from a large pothole on Burnt

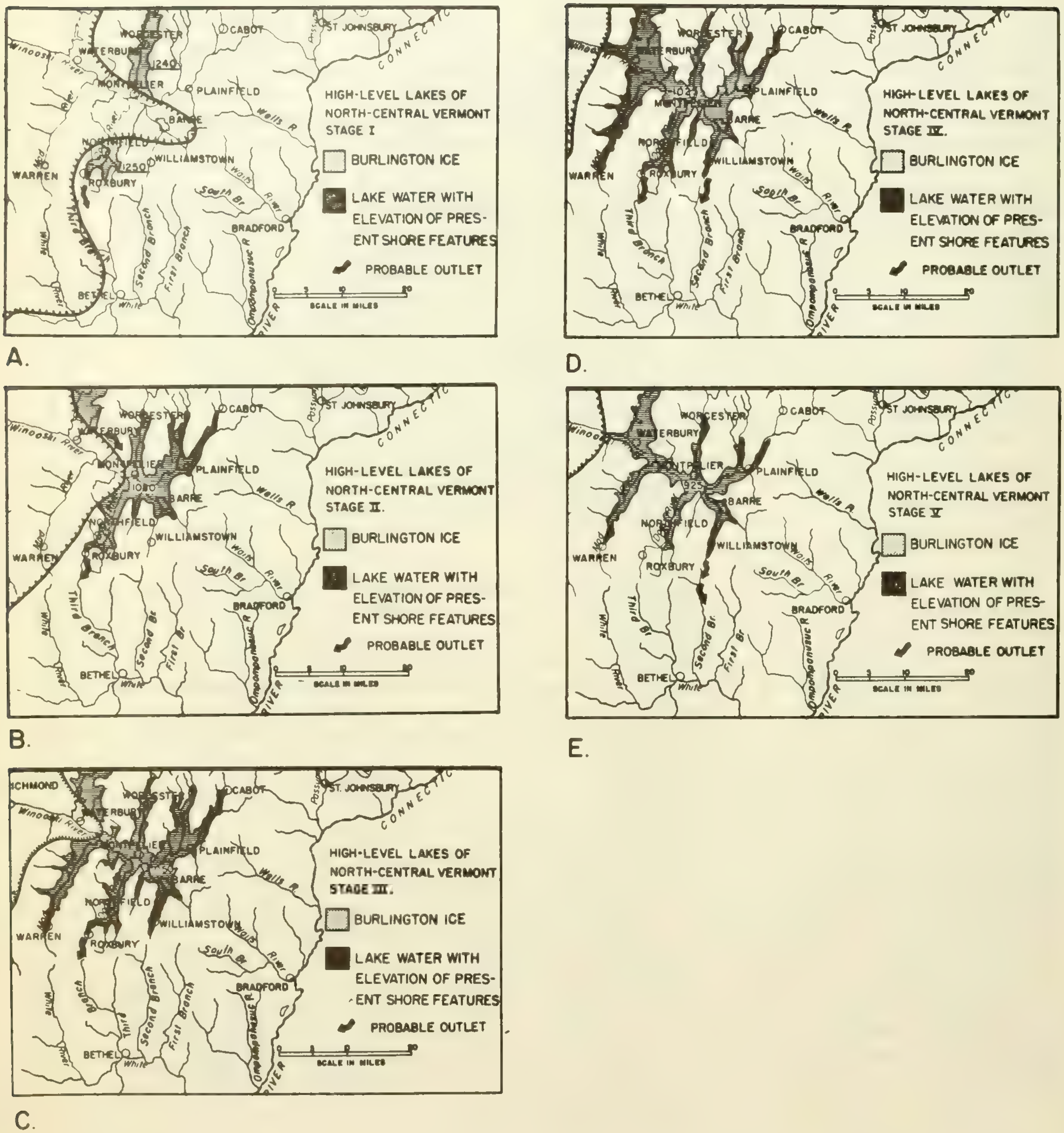


Figure 4. High-level lakes of central Vermont according to Stewart and MacClintock (1969, figs 18-22, published by permission of Dr. Charles G. Doll, Vermont State Geologist). Only the lower half of the original figures were reproduced.

Rock Mountain situated on the crest of the Green Mountains 16 miles due west of Montpelier. The pothole, described by Doll (1936), is at an approximate elevation of 2820 feet. Continued downwasting resulted in long coalescent masses of stagnant ice filling the valleys of the Winooski River and its tributaries. Drainage in the main Winooski valley was blocked, therefore the surfaces of the ice masses probably rose to the northwest with a low gradient.

A sequence of early proglacial lakes that formed in the central Vermont area, according to Stewart and MacClintock (1969), is shown in figure 4. The sequence of diagrams clearly implies that thresholds at Roxbury and south of Williamstown are erosional, having been lowered 240 feet and 110 feet respectively. It is the contention of this report that neither threshold was affected appreciably by runoff from glacial lakes (possibly 5 to 20 feet of till were removed from each threshold), because ice-contact features immediately north of each threshold are constructional in origin.

Proglacial lakes developed where north-flowing tributaries, such as the Mad River, the Dog River, and the Stevens Branch, were dammed on the north by stagnant ice (further discussion of the Mad River is not included in this report). These lakes drained southward over bedrock thresholds into the drainage system of the Connecticut River (fig. 5).

In the Dog River valley there are four groups of terrace levels, (1) 1010 to 1020 feet, (2) 910 to 920 feet, (3) 740 to 760 feet, and (4) 640-680 feet, which punctuate the history of deglaciation into four stages. The first three groups of terraces consist mostly of constructional surfaces (deltas, kame deltas or kame terraces), and were controlled by proglacial lakes, the sequence of which depended upon the position of an ice margin during deglaciation. The fourth group of terraces is believed to be mostly erosional, as are terraces whose elevations do not fall within one of the four major groups

Stage I

Glacial Lake Roxbury: The highest terraces are associated with a lake which was controlled by a threshold at 1010 feet elevation at Roxbury, and which drained southward by way of the Third Branch of the White River (fig. 5). This lake was first noted, but not named, by Merwin (1908, p. 124). It is named here glacial Lake Roxbury. The major evidence for a lake at 1010 feet is a large ice-contact delta (Stop 9) situated 1.3 miles north-northeast of Roxbury. Foreset bedding, ripple-drift cross-lamination, and dune bedding, each indicating a southward transport direction, are exposed in a sand and gravel pit, now used for a sanitary landfill. The contact between topset and foreset bedding at 1012$\pm$ feet elevation is exposed in the southwest corner of the pit. The delta, 0.8 of a mile long, is a constructional feature since its surface is pock-

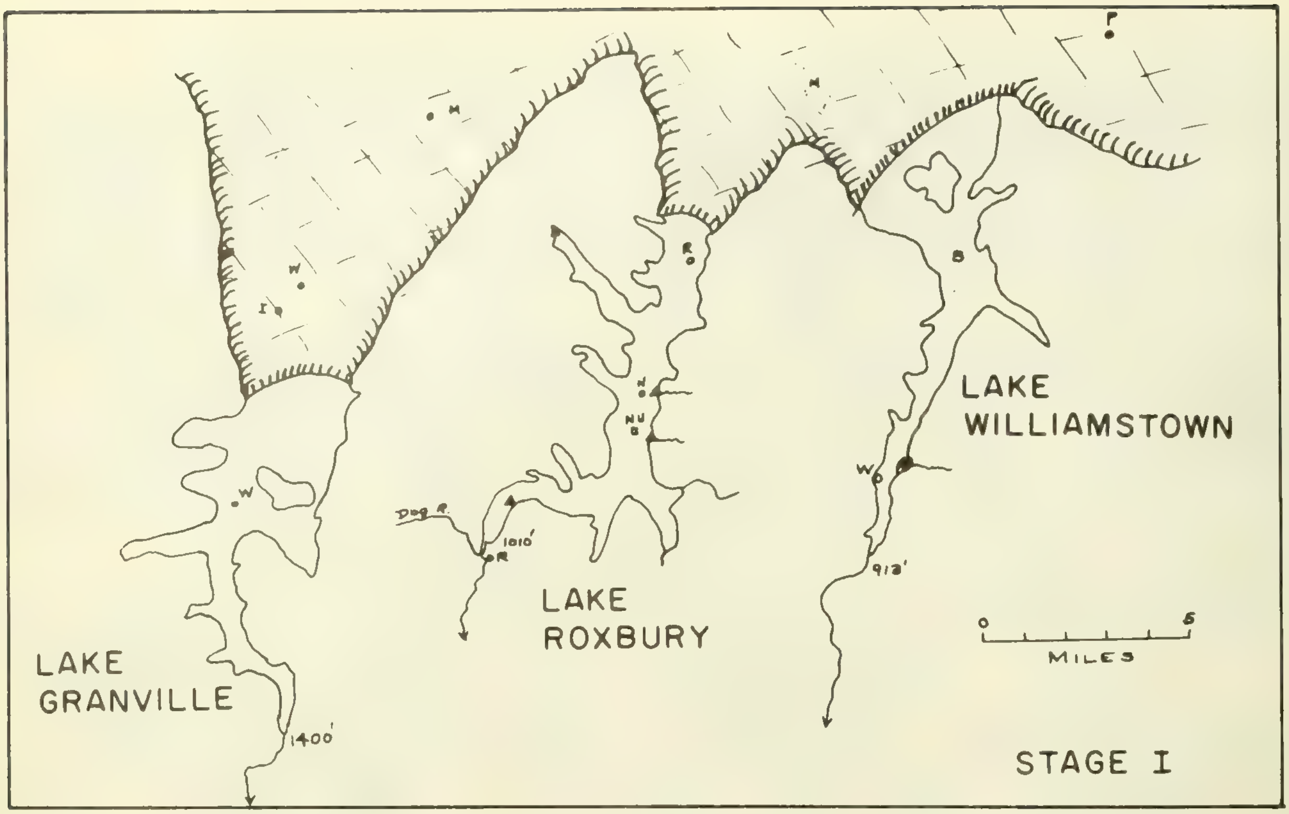


Figure 5. Stage I, Lakes Williamstown, Roxbury, and Granville.

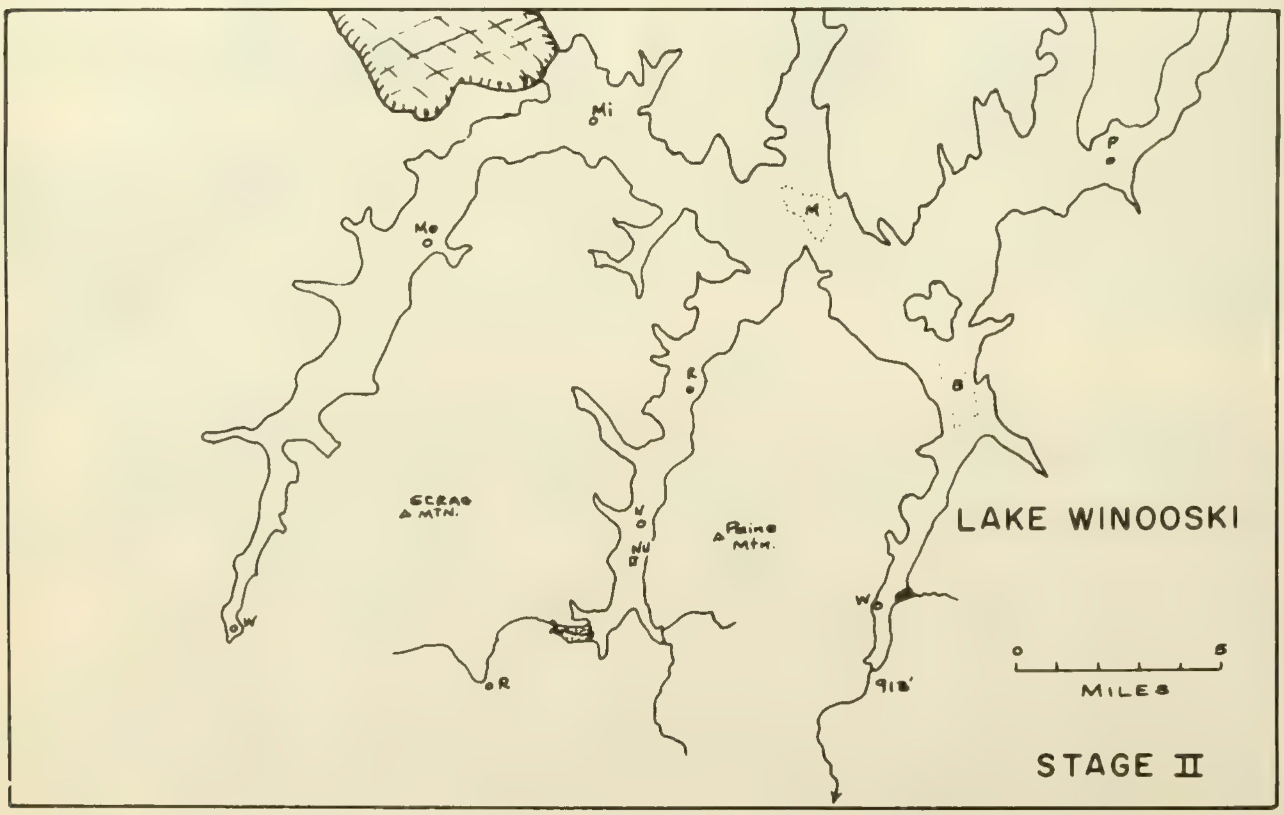


Figure 6. Stage II, Lake Winooski.

marked with kettles on the north, and it was fed by a subglacial stream as indicated by an esker which extends 1.2 miles east-southeast from the head of the delta.

Retreat of the ice margin from the ice-contact deltas was accompanied by the northeastward expansion of Lake Roxbury. Just how far north the 1010-foot lake extended is not known, however features in ice-contact gravels 1.0 miles north of Northfield indicate a transport direction to the south. Also ripple-drift cross-lamination in lacustrine sands at Riverton dips to the south. Three small 1000-foot terraces (kame deltas) on the east side of the Dog River valley at Northfield may or may not have been deposited in Lake Roxbury. The location of the three features in relation to the post office at Northfield is as follows: (1) 0.3 of a mile east, (2) 0.75 of a mile south, and (3) 1.4 miles south. Good exposures are lacking in the three deltas, therefore direction of transport and topset-foreset relationships are unknown.

Glacial Lake Williamstown: Shortly after the formation of Lake Roxbury, a proglacial lake developed in the valley of the Stevens Branch. The lake, named Lake Williamstown by Merwin (1908, pl. 21B), drained over a threshold at 915 feet elevation, 2.3 miles south-southwest of Williamstown (fig. 5). Southward dipping foreset beds in a kame terrace, 0.25 of a mile east of Williamstown, clearly indicate the former presence of a standing body of water. Drainage of the lake was to the south through Williamstown Gulf by way of the Second Branch of the White River. That stagnant masses of ice choked the Stevens Branch valley during deglaciation is shown by the plentiful occurrence of eskers and kame terraces for at least 5 miles north of the threshold. The presence of unfilled kettles in the kame terraces testifies to the constructional origin of the land forms.

Continued downwasting and retreat of the ice margins bordering Lakes Roxbury and Williamstown finally resulted in the lowering of Lake Roxbury by 95 feet to the level of Lake Williamstown. This occurred when ice withdrew below the 1000 foot contour (approximate) on the ridge separating the valleys of the Dog River and the Stevens Branch. The locality is on the Barre quadrangle, 2.5 miles north of Berlin.

Stage II

Glacial Lake Winooski: The second group of terraces in the Dog River valley, at 910 to 920 feet, lies 90 to 100 feet below the former level of Lake Roxbury. The features are best developed in the vicinity of Harlow Bridge School, 2.25 miles south-southwest of Northfield. A large delta, with surface elevations greater than 920 feet, is situated 0.6 of a mile west of Harlow Bridge School. Well developed terraces lie above the 900 foot contour northwest, southwest, and south of Harlow Bridge School.

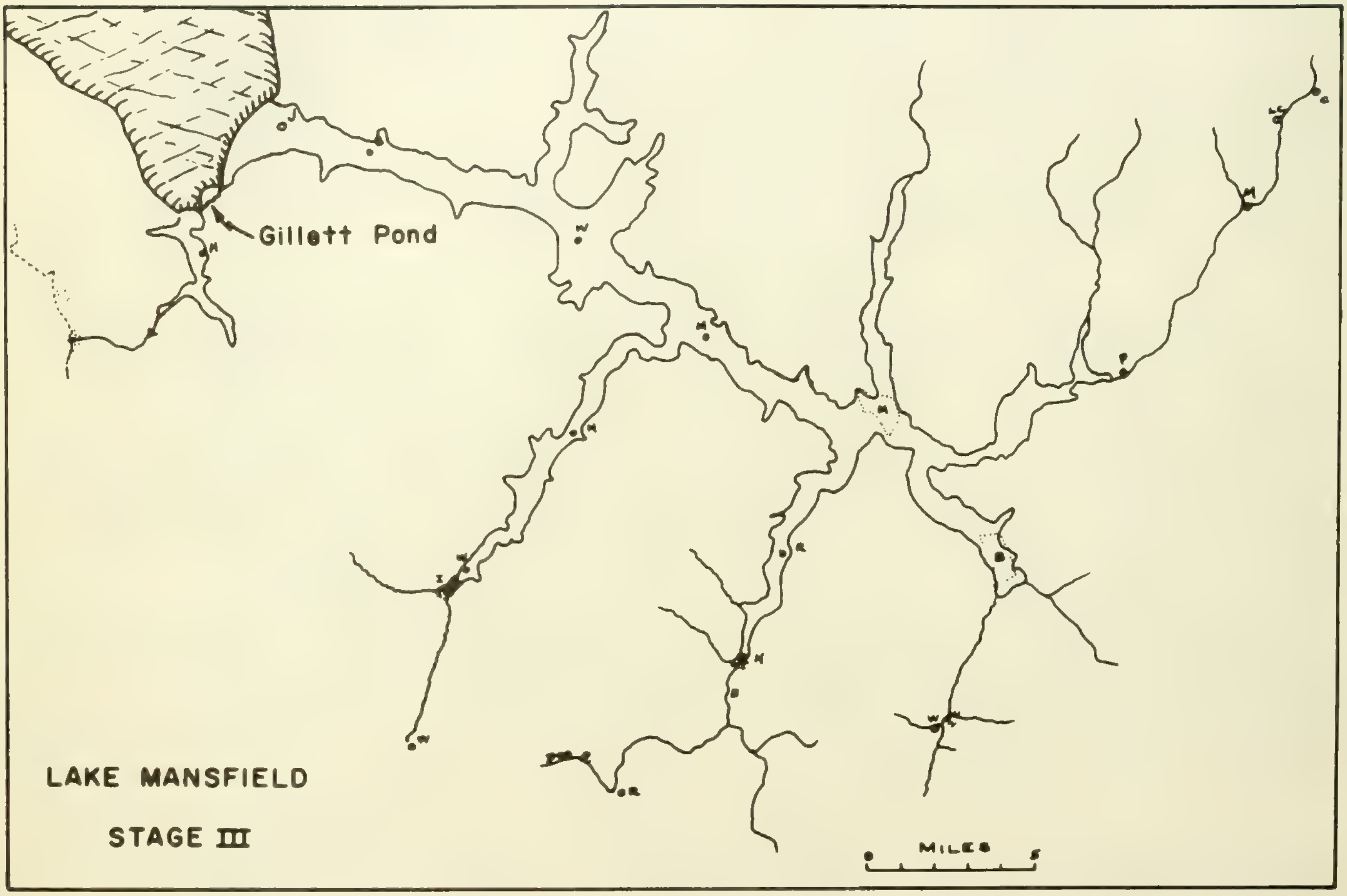


Figure 7. Stage III, Lake Mansfield. This map is intended as a first approximation only, and was obtained by tracing the 700-foot contour on the Lake Champlain sheet of the AMS 1:250,000 series.

Foreset beds of fine sand and silt occur in a small pit 0.1 of a mile southeast of the Harlow Bridge School. Here, ripple-drift cross-lamination at 850 feet elevation indicates that during deposition transport of sediment was to the north. An area of hummocky ground with summits at 900 feet elevation extends for 1.2 miles south of Norwich University. The area has a core of bedrock and ice-contact stratified drift, the latter displaying features indicating a southward transport direction. Covering the bedrock and drift core is a mantle of sand and silt. The area probably was underlain by masses of buried ice which were covered by deltaic and lacustrine sediments deposited by Sunny Brook which enters the Dog River 1.8 miles south-southwest of Northfield. Melting of the ice blocks resulted in the collapse of the 900 foot delta surface. Similar collapsed topography occurs in the vicinity of the former Northfield dump, 1.0 mile north of Northfield.

The development of terraces and deltaic surfaces at 910 feet elevation requires the presence of a lake at that approximate elevation. This lake is glacial Lake Winooski (fig. 6), which was formed by the coalescence of Lake Roxbury and Lake Williamstown. The development of a major new lake by the coalescence of two previously named lakes is assumed here to require a different name for the single lake thus formed. Merwin (1908, p. 138) used the term "First Lake Winooski" to describe a lake which was blocked by an ice margin between Middlesex and Plainfield and lower portions of the valleys of the Dog River and the Stevens Branch. However, First Lake Winooski was "represented by an altitude of 745 feet at Plainfield" (Merwin, 1908). It is not clear where the outlet of the lake was situated, however, Merwin must have assumed that it was over ice toward the west-northwest. In the stage following First Lake Winooski, Merwin shows Lake Mansfield with an outlet along the ice margin west of the Green Mountains. Because Lake Mansfield may be a valid term, and because the outlet of First Lake Winooski is questionable, the term Lake Winooski is used to describe the temporary proglacial lake which drained south through Williamstown Gap after the coalescence of Lake Roxbury and Lake Williamstown.

Stage III

Lake Mansfield: In the Dog River valley there is a wide range of constructional and erosional terraces below 900 feet elevation. However, the next most consistent group of terraces occurs between 740 and 760 feet. Union Brook, Cox Brook, and Chase Brook, southeast-flowing tributaries of the Dog River, have each built small deltas into a lake at this level at Northfield, Northfield Falls, and just north of Riverton, respectively. Delta surfaces are common between 720 and 760 feet elevation throughout the upper Winooski drainage area suggesting that they all share a common origin in a single lake. Since the lowest divide between the Champlain valley and the Connec-

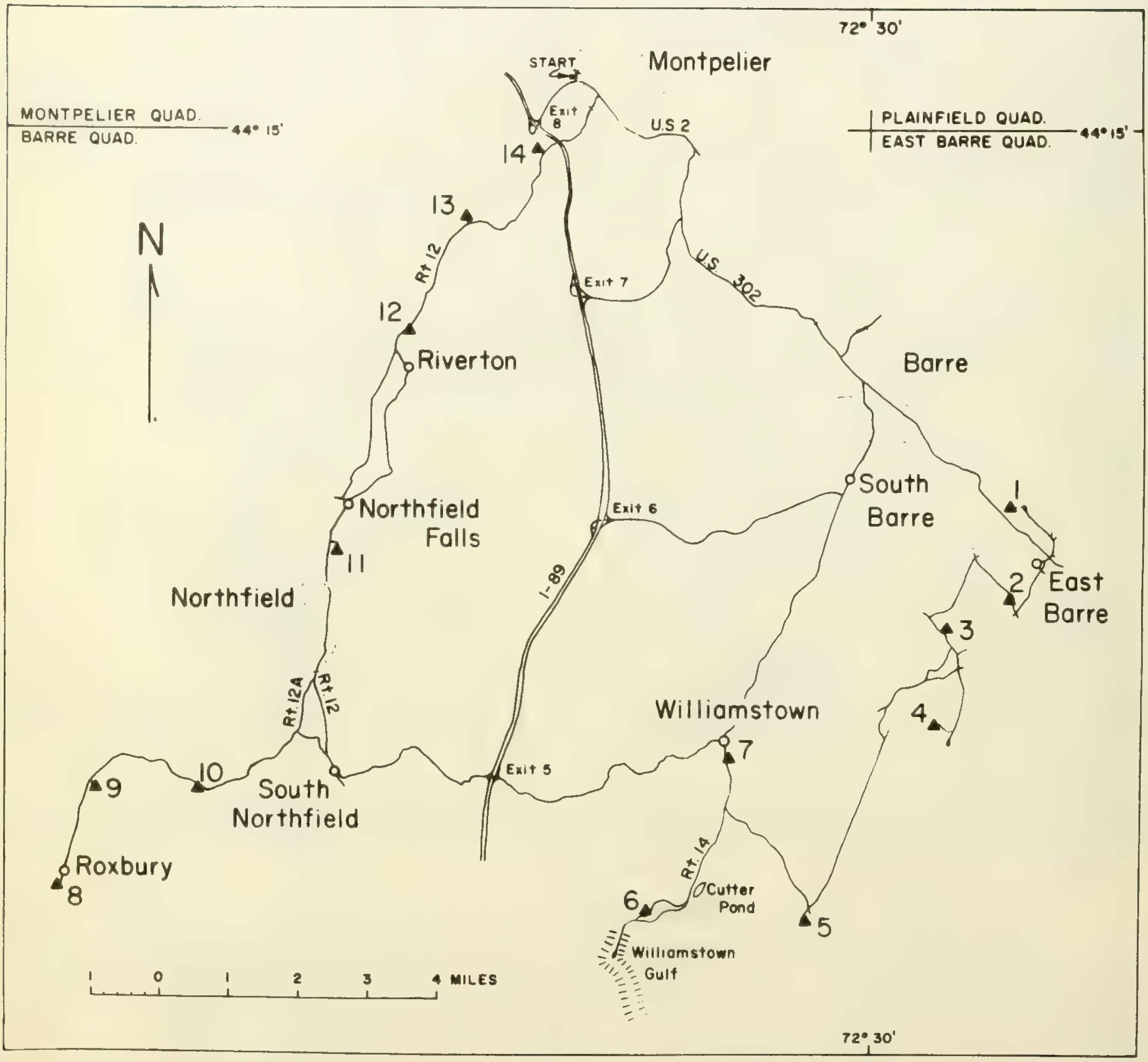


Figure 8. Route for Field Trip G-1. Solid triangles denote stops.

ticut valley is the 915-foot threshold south of Williamstown, the outlet for a lower lake must lie west of the Green Mountains.

Merwin (1908) suggested the name Lake Mansfield (fig. 7) for a lake in the Winooski valley which had an outlet along an ice margin in the vicinity of Huntington. A possible outlet for Lake Mansfield lies just northeast of Gillett Pond, 2.9 miles N 31° E of Huntington. However, the Gillett Pond threshold lies at an elevation of 740 feet which is the same elevation as the terraces in the Dog River valley. Studies in southern Quebec (McDonald, 1967), the Champlain valley (Chapman, 1937), and the lower Connecticut valley (Jahns & Willard, 1942) indicate that regional tilting of the surface of New England has occurred since removal of the weight of the last continental ice sheet. The amount of tilt that has occurred since late-glacial time is on the order of 4 feet per mile. Since Gillett Pond lies 20 miles northwest of the Dog River valley (measured perpendicular to isobases), the outlet should be approximately 80 feet higher than the 740-foot terraces in the Dog River valley. Since the Gillett Pond threshold lies at 740 feet (approximate), either (1) the outlet has been lowered 80 feet by erosion, or (2) the 740-foot terraces in the Dog River valley were deposited in higher lakes controlled by temporary thresholds related to blocks of stagnant ice, or (3) some combination of these two has occurred.

Stage IV

Well-formed terraces occur at elevations of 640 to 680 feet in the Dog River valley. Since these features are erosional and slope down valley, they can probably be related to one of two possible situations. The first is a glacial lake with a 660-foot threshold through Hollow Brook, 2.0 miles S 36° W of Huntington. Diversion of drainage over the Hollow Brook threshold would require blockage of the lower Winooski valley in the vicinity of Richmond following retreat of ice from the Gillett Pond outlet. If such a blockage did not control Stage IV terraces in the Dog River valley, then possibly they are graded to a level of Glacial Lake Vermont. Since close field inspection has not been made of possible thresholds west of the Green Mountains, the above discussion of Stages III and IV must be considered conjectural at this time.

Road Log

Mileage

START: MONTPELIER QUADRANGLE

0.0 Begin mileage count and turn right at intersection of Bailey Avenue (U.S. 2) and exit from Montpelier High School parking lot. The parking lot is on the flood plain of the Winooski River and was under 6 feet of water during the flood of 1927.

0.1 Cross railroad tracks and turn left on Memorial Drive, proceed east on U.S. 2 along the south bank of the Winooski River.
Continue straight ahead at traffic light, crossing Rt. 12.
0.6 Striated exposure of Waits River Formation on the right (see fig. 2)
1.05 BARRE QUADRANGLE
2.2 Turn right (south) and follow U.S. 302 to East Barre. Route leaves the Winooski River and follows the valley of the Stevens Branch.
3.0 Varved clay on the right represents bottom sediments of glacial Lake Winooski, or glacial Lake Mansfield. In the spring of 1960, this locality was the site of an earthflow which covered 2 of the 3 lanes of the Barre-Montpelier Road.
3.3 Material and Soils Laboratory of the Vermont State Highway Department on the left.
6.6 Rt. 14 enters from the left, continue straight ahead.
7.1 Bear right, then turn left around municipal park in the center of Barre, follow U.S. 302. Road ascends to 720 foot terrace (delta?).
7.2 EAST BARRE QUADRANGLE
7.5 Exposure behind gas station on the left has lacustrine sand which contains angular ice-rafted pebbles, and which is overlain by till. Road enters the valley of the Stevens Branch.
8.6 View ahead of Cobble Hill.
9.3 View of Jail Branch section on left.
10.4 Turn left at junction of U.S. 302 and Rt. 110
10.5 Bear left at Y.
10.7 Bear left as 2 roads branch to the right.
11.3 Park on right at gravel pit in ice-contact stratified drift. Cross road and walk southwestward across field to top of bank.

STOP 1. Jail Branch Section. From the base up, the section consists essentially of fine-grained lacustrine sediments (silt and clay) which grade upward into lacustrine fine sand and silt which, in turn, is overlain by gravel and sands of probable outwash origin, and finally till with large erratics of Barre Granite. This sequence is believed to be the result of blockage of the Jail Branch by advancing ice which finally encroached upon and overran a lacustrine sequence. The sequence is preserved because it is situated in the erosional shadow of Cobble Hill.

Retrace route to U.S. 302.
12.2 Continue straight ahead (south) crossing U.S. 302 and the Jail Branch.
12.3 Turn left at Y, follow Rt. 110.
12.4 Bear right, leaving Rt. 110, on road to Upper Graniteville.
13.1 Turn right (west) on dirt road which passes several large grout sites.
13.4 Park on right side of road.

STOP 2. A brief photographic stop at abandoned quarry to view sheeting in the Barre granite.

13.9 View of Jail Branch section and Cobble Hill to the right.
14.2 Stop sign, turn left (south).
15.4 Turn left (southeast).
15.6 Rock of Ages Tourist Center on the right.
15.7 Park on right side of road adjacent to Rock of Ages quarry.

STOP 3. Gray till overlies Barre granite and basic dike. The till contains numerous striated clasts of calcareous quartzite derived from the Waits River Formation. Striae on granite trend due south.

Proceed straight ahead, roads enter from the right, then from the left.
16.3 Continue straight (south) at crossroad.
17.4 Park on right side of road, walk to top of Mount Pleasant, elevation 2063'.

STOP 4. Barre Granite Indicator Fan and Boulder Train. Mount Pleasant is underlain by gray phyllite and slate of the Gile Mountain Formation. However, the top is covered with numerous granite erratics of medium-grained, gray granite. The nearest outcrop of Barre granite lies 0.6 miles to the north and 500 feet lower than the summit of Mt. Pleasant.

Return to cars, proceed straight ahead.
17.5 Make U-turn in driveway of summer home.
18.6 Turn left (west) at crossroad.
19.1 Continue straight ahead with caution, road enters from the right.
19.9 Turn left (south) to Baptist Street.
20.4 Bear right as route enters Baptist Street.
20.9 BARRE QUADRANGLE
21.1 View to the west of Paine Mountain and the Green Mountains.
23.1 Jackson Corner, continue straight ahead.
23.3 Park on right side of road.

STOP 5. White Rock. A mass of vein quartz measuring 115 x 45 x 15 feet is the crag of a large-scale crag-and-tail feature. The axis of the tail trends due south supporting the contention that the last important ice movement in this area was due south. Numerous blocks of vein quartz may be found located in a stone wall 375 feet south of White Rock.

Return to cars, reverse direction either by backing up or by proceeding south to farm at end of road.
23.5 Turn left at Jackson Corner, road descends into the valley of the Stevens Branch.
25.5 Turn left (south) on Rt. 14, gravel pit on west side of valley is in ice-contact stratified drift graded southward to the threshold of glacial Lake Williamstown.
26.3 Esker on the left.

26.7 Cutter Pond, elevation 912 feet on the left.
26.8 Threshold of glacial Lake Williamstown, approximate elevation 915 feet. Road descends into Williamstown Gulf, a V-shaped valley deepened by the outlet from glacial Lake Williamstown.
28.3 Turn to the right into parking lot of restaurant in Williamstown Gulf, make U-turn with caution and rejoin Rt. 14 north.
28.8 Turn left on dirt road.
30.0 Stop at Staples Pond, elevation 890 feet.

STOP 6. Outlet of Lake Williamstown and Lake Winooski. During deglaciation, when drainage to the north was blocked by stagnant ice, Lake Williamstown formed north of Cutter Pond and drained southward through this area to Williamstown Gulf.

30.9 Turn left (north) on Rt. 14, proceed to Williamstown.
33.1 Turn right into yard of Burrell Roofing Compamy

STOP 7. Ice-contact features at Williamstown. The sheet metal shop is located at the south end of a discontinuous esker which shows on the Barre quadrangle as a single closed contour at 880 feet elevation. Foreset beds and ripple-drift cross-lamination dip to the south. Collapsed and faulted beds are common behind the sheet metal shop and in a small pit 100 feet to the north. The view to the northeast is of a partially excavated kame delta with foreset beds which dip to the south. In view of the widespread occurrence of ice-contact features deposited in relation to the 915-foot threshold of Lake Williamstown, the topography is considered to be constructional.

Proceed north on Rt. 14.
33.4 Turn left (west) leaving Rt. 14 at center of Williamstown.
34.6 Bear right at Y.
35.6 View left (east) into the valley of the Stevens Branch.
36.1 Drainage divide, elev. 1715 feet, enter drainage basin of Dog River.
37.5 Pass under I-89, several sharp curves ahead.
40.2 Stop sign, turn right (northeast) on Rt. 12 in the village of South Northfield (!) situated in the valley of Sunny Brook.
40.5 Turn left (west) on dirt road which leaves Rt. 12 and follows Sunny Brook.
41.1 Stop sign, turn left (south), follow Rt. 12A and Dog River valley to Roxbury.
46.3 Turn right to railroad depot at Roxbury.

STOP 8. The depot at Roxbury is situated on the drainage divide of a small through valley. The Dog River descends from the slope on the west and turns to the north, whereas the Third Branch of the White River enters the valley from the east and turns to the south. The drainage divide, at an elevation of 1010 feet, is the former threshold of glacial Lake Roxbury.

Proceed north on Rt. 12A.

46.8 Camp Teela-Wooket on the right.
47.0 Rt. 12 passes through terrace graded southward to the Roxbury threshold.
47.2 Terrace on the right.
47.6 Rt. 12 rises on the front of ice-contact delta.
47.8 Turn right into gravel pit being used as sanitary landfill dump.

STOP 9. Ice contact delta. Foreset bedding, dune bedding, ripple-drift cross-lamination, and imbricate structure indicate southward transport of sediment during construction of the delta. Maximum height of foreset beds overlain by topset beds is on the order of 1012 to 1015 feet indicating deposition in a lake whose elevation was controlled by the Roxbury threshold. Collapsed bedding, kettles, and an esker, which extends 1.2 miles down the Dog River valley, give evidence of an ice-contact origin for the delta. Headward erosion increased the length of the gully at the southeast corner of the pit by 50 feet between October, 1970, and October 1971.

Return to Rt. 12A, proceed north.
48.7 Railroad overpass and bridge over Dog River.
50.1 Park on right side of Rt. 12A, cross wooden bridge over Dog River, enter pit.

STOP 10. Neun Pit (tentative stop). Sediments in the lower portion of pit are gravel, sand, and silt which display foreset beds (bar slip faces?) and dune bedding which indicate transport of sediment to the west, or up the Dog River valley. Transport direction in the overlying stream gravel was to the east as shown by imbrication of pebbles. The lower sediments are assumed to be ice-contact deposits formed by a subglacial stream flowing into Lake Roxbury. The upper stream gravels were deposited by the Dog River which, at the time of deposition, was graded to Lake Mansfield or to later stage deposits.

Proceed north on Rt. 12A.
50.5 Cross Dog River in middle of Northfield Country Club. Skyline to the left (north) is the surface of a 920-foot delta formed in Lake Winooski. Note terraces on the golf course at the right.
51.3 Pass under Harlow Bridge, scene of famous 1867 railroad disaster in which several railroad cars were accidently pushed from half-completed bridge.
51.4 Harlow Bridge School on left. Pit to the right on Bull Run Road has fine sand in bottomset beds, or low-dipping foreset beds, deposited in Lake Winooski. Ripple-drift cross-lamination at 850 feet elevation dips to the north.
51.9 Bridge over Sunny Brook. For the next 0.8 of a mile hummocky ground lies on the right.
52.7 Stop sign, turn left (north) on Rt. 12.
52.8 Turn left at small park.
52.9 Park on right for brief rest stop at Norwich University. Proceed north on Rt. 12 through the village of Northfield.

53.7 Downtown Northfield (Depot Square).
53.8 Bridge over Dog River.
53.9 Traffic light, terrace to left is surface of 740-foot delta built into Lake Mansfield.
54.6 Bridge over Dog River.
54.7 Turn right (east) on dirt road just past Catholic Cemetery. Road rises to 700-foot terrace.
54.9 Park on right, walk south along 700-foot terrace to former site of Northfield town dump.

STOP 11. Collapsed lacustrine sediments are exposed in a face 200 feet long and 15 to 30 feet high. Thick layers of fine sand and silt at the base grade upward into thin layers of varved silt and clay. Ripple-drift cross-lamination in the sand layers dips to the north. Angular, ice-rafted clasts of fine-grained chlorite schist and greenstone occur in a layer about 10 feet above the base of the section. Striations occur on a greenstone clast which measures 2' x 2' x 1'. The occurrence of fine-grained lake-bottom sediments at elevations up to 760 feet suggests deposition in glacial Lake Winooski (Stage II). The presence of angular clasts testifies to the presence of icebergs in the lake,and large scale collapse,as shown by dipping and faulted beds,indicates lacustrine sedimentation over buried ice. Large folds formed by collapse were once exposed in a lower portion of the pit now covered by the dump. Gravel overlies collapsed and truncated layers of fine sand at the right.

55.1 Turn right (north) on Rt. 12.
55.2 In the pit at the right ice-contact gravels with features indicating southward transport capped by stream gravels with imbrication suggesting northward transport. The stream gravels are the same deposits that underlie the 700-foot terrace at STOP 11.
55.8 Turn left (west) at IGA Store in Northfield Falls, continue through covered bridge over Dog River and over railroad tracks.
55.9 Turn right (north) just beyond railroad tracks.
56.3 Terrace at 660 feet elevation underlies red barn on the right. Road ascends bedrock spur.
56.5 Brief photographic stop on the right, time and weather permitting.
56.7 Road descends to 680-foot terrace with view of three erosional terraces below. Exposure to the right is in ice-contact gravels with directional features oriented to the south.
57.1 Road drops to 660-foot terrace.
57.3 Road drops to brook and reascends to 660-foot terrace.
57.9 Entrance to gravel pits on the right.
58.1 Turn right into driveway which circles West Berlin (Riverton) School and park.

STOP 12. Riverton Water Gap. During Stage III a 740-foot delta was apparently deposited across the preglacial course of the Dog River, 0.6 of a mile due north of the West Berlin School. Lowering of Lake Mansfield, caused by a change in outlets west of the Green Mountains, permitted the superposition of the postglacial Dog River across a bedrock spur 0.7 of a mile north-northeast of West Berlin School.

Proceed north on dirt road.
58.3 Stop sign, bear left (north) on Rt. 12.
58.9 Riverton water gap. Bedrock exposed is fine-grained chlorite schist of the Cram Hill Member of the Missisquoi Formation.
60.0 Slump terracettes on the left.
60.5 Turn left (northwest) on road to pit.

STOP 13. Herring Pit (tentative stop). Ice-contact stratified drift with features indicating southward transport occur at elevations up to 680 feet.

Proceed north on Rt. 12.
61.3 Abandoned potholes occur in a railroad cut behind a trailer on the left. They appear to have been cut by the Dog River before the last glacial advance because they probably were filled with lacustrine sediment that occurs adjacent to the railroad cut.
61.5 Road follows flood plain of the Dog River for 1.1 miles.
62.5 Turn left (northwest) on dirt road.
62.6 Turn left (southwest) to exposure.

STOP 14. Collapsed mass of ice-contact stratified drift. The exposure is all that remains of a small hill 0.4 of a mile southeast of the point where the Dog River leaves the Barre quadrangle. The original feature was 500 feet long and over 50 feet high. All types of glacial sediments, including till, have been seen in the hill as it was being reduced by man. Highly distorted clay-silt varves presently overlie boulder gravel on a contact that strikes N 15° E and dips 40° northwest. Is the feature a constructional or an erosional land form?

Return to Rt. 12, proceed north.
63.1 Enter Montpelier
63.4 MONTPELIER QUADRANGLE
64.0 Traffic light, turn left (west) on U.S. 2.
64.4 Turn right over railroad tracks.
64.5 Turn left into parking lot of Montpelier High School.

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"We are unable to adopt these views; first, because all known glaciers are confined to valleys, though at their head they may be connected with extensive fields of ice, capping the summits of the mountains: secondly, because no known glacier is more than 50 or 60 miles wide (the great glacier called Humbolt, in Greenland, described by Dr. Kane, is of this width), whereas the ancient American glacier must have been at least 2500 miles wide, and have spread over all the mountains as well as valleys, and often have been obliged to move up hill as well as over a level surface: thirdly, because in our country we have two and probably three prominent directions to our drift, and it is difficult to see how one glacier would have moved in so many directions, especially as the most usual course of the striae in New England does not follow a valley, but crosses over mountains obliquely."

. . . .Edward Hitchcock, 1861
Geology of Vermont, v. 1, p. 91.

Trip G-2

ICE MARGINS AND WATER LEVELS IN NORTHWESTERN VERMONT

by

W. Philip Wagner
University of Vermont

PROGLACIAL LAKES IN THE LAMOILLE VALLEY, VERMONT

by

G. Gordon Connally
State University of New York at Buffalo

ICE MARGINS AND WATER LEVELS IN NORTHWESTERN VERMONT

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INTRODUCTION

In what has become a classic reference for late Pleistocene drainage history in the Champlain Valley, Chapman (1937) delineated a series of lacustrine and marine water bodies associated with retreat of the Laurentide ice sheet. Successively lower levels of proglacial Lake Vermont extended progressively further northward, following the retreating ice margin. Finally, ice retreat allowed the influx of marine waters forming the Champlain Sea (Karrow, 1961). Numerous investigators working in the Green Mountain uplands have recognized the existence of local lakes, which were impounded between the highly irregular topography and the Laurentide ice margin, and which were partly contemporaneous with Lake Vermont. The publications by Connally (1966) and Stewart and MacClintock (1969, 1970) are recent examples.

This report summarizes research on Pleistocene proglacial events in the Champlain Valley and adjacent Green Mountain uplands. Numerous students at the University of Vermont provided assistance, including R. Switzer, C. A. Howard, Jr., W. R. Parrott, Jr., and B. P. Sargent. The use of data from dissertations by Johnson (1970) and Waite (1971) is gratefully acknowledged. G. G. Connally, C. S. Denny, and B. C. McDonald reviewed early drafts of the manuscript. The work upon which this report is based was supported by funds provided by the United States Department of Interior as authorized under the Water Resources Research Act of 1964, Public Law 88-379.

WATER PLANES

General

Raised strandlines in the northern part of the Champlain Valley are marked by abundant but widely scattered shoreline features consisting primarily of deltas and beaches, but also including outlet channels, wave-cut benches, and spits. The locations of these features are shown in Figures 1 and 2. A listing of features, with pertinent information is provided in the appendix. Figure 3 is a north-south profile, constructed by westerly projection of features, with elevation control provided by contour lines from topographic maps.

Delineation of water planes is difficult in this area due to

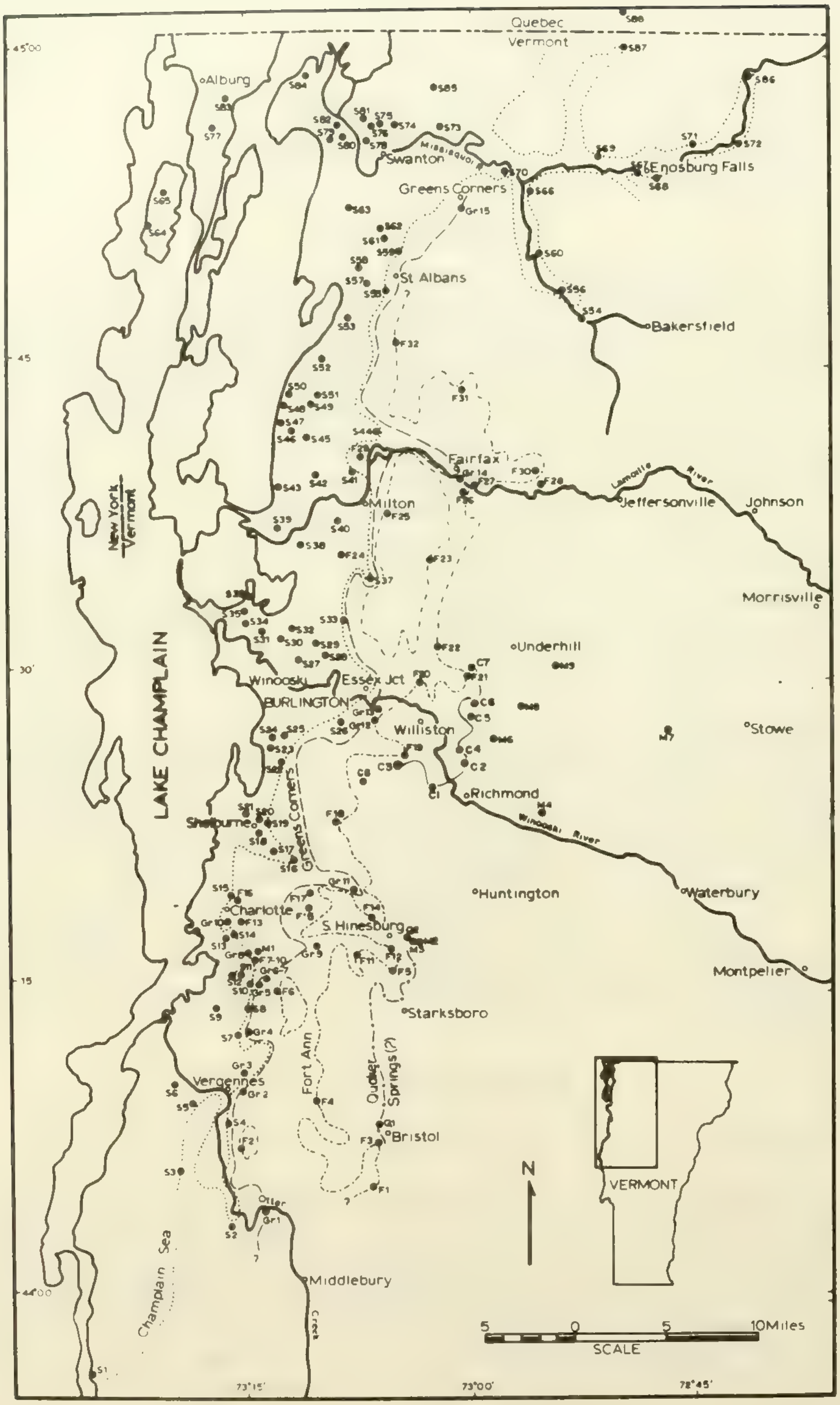

Figure 1: Shoreline feature locations and strandlines of regional water bodies in Champlain Valley: S = Champlain Sea; Gr = Greens Corners; F = Fort Ann; C = Coveville(?); Q = Quaker Springs(?); M = Miscellaneous.

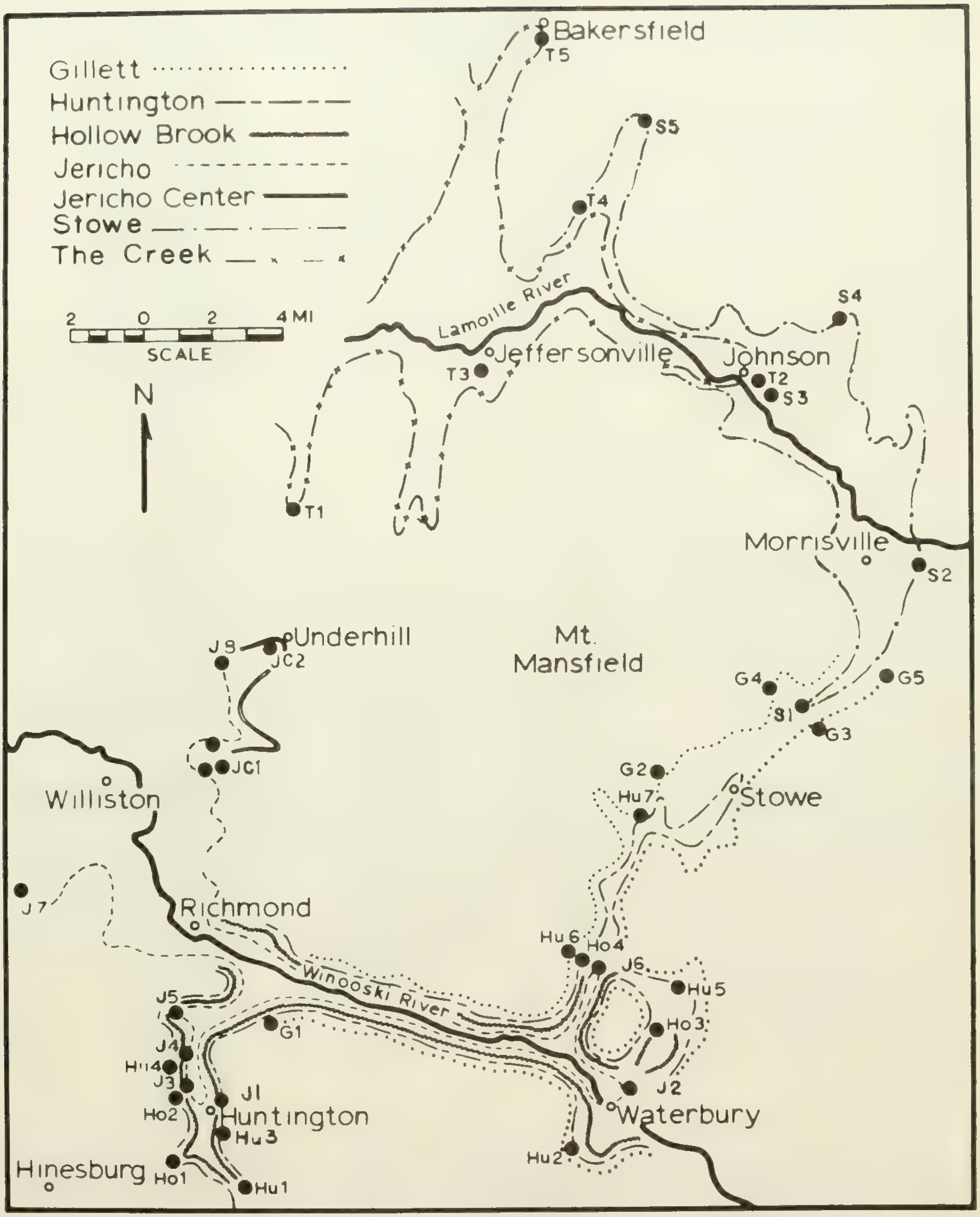

Figure 2: Shoreline feature locations and generalized strandlines of upland water bodies in the Green Mountains: G = Gillett; S = Stowe; T = The Creek; Hu = Huntington; Ho = Hollow Brook; J = Jericho; Jc = Jericho Center.

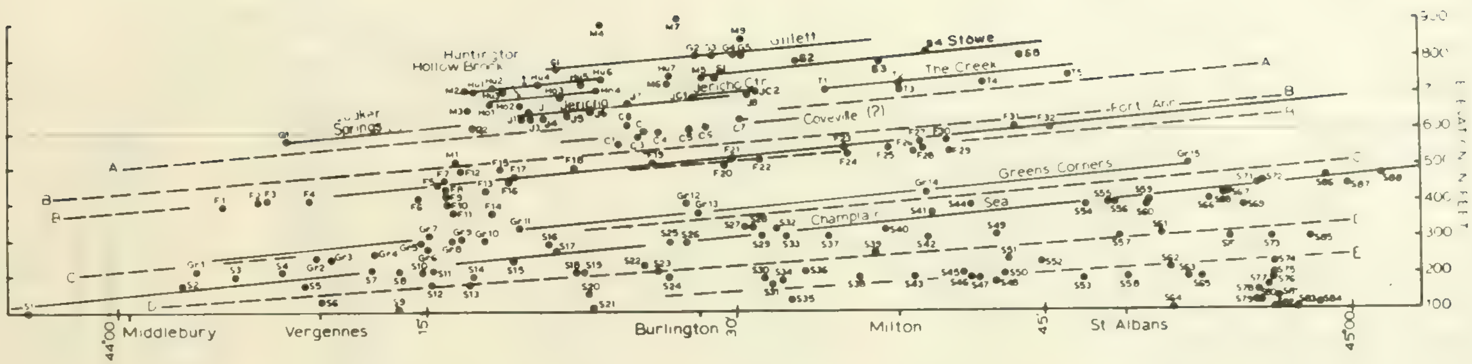

Figure 3: Water planes of regional and local water bodies on a north-south section. Letter symbols are same as for Figures 1 and 2. Dashed lines A, B, C, D and E approximate Chapman's (1937, Figures 15 and 16) Coveville, Fort Ann, upper marine, Port Kent and Burlington planes, respectively.

the large scatter of shoreline features. The most obvious alignment of features on Figure 3 approximates the marine limit (Champlain Sea) of Chapman (1937, Figure 16), which is different from the marine limit of this paper based on the highest occurrences of marine fossils (Figure 3; Appendix). The Fort Ann (Chapman, 1937) and Greens Corners water planes on Figure 3 are drawn parallel to the marine limit so as to coincide with both the largest number of features possible as well as the more prominent features. Above the Fort Ann level distinct regional water planes are not apparent. The Coveville (Chapman, 1937) and Quaker Springs (Stewart, 1961) planes are tentatively recognized, based on correlation with features identified by others (Connally, 1968 and 1970; Denny, 1970, personal communication). Features above the Quaker Springs level represent local lakes in the Green Mountains. By considering topography, distribution of shoreline features, drainage requirements, and assumed configurations of the Laurentide ice margin, water planes for local lakes above the Quaker Springs level were drawn parallel to the regional, lowland water planes.

The accuracy of Figure 3 is affected by a variety of sources of error. If combined, errors could result in some features being misplotted 40-50 feet too high or low on Figure 3. Comparison of Figure 3 with a similar profile from the New York side of the Champlain Valley by Denny (1970, personal communication) indicates very close agreement for the major regional strandlines common to both profiles (marine limit; Fort Ann; Coveville[?]). Water planes for local, upland lakes are considered tentative in view of data limitations.

Existing terminology has been considered in naming the various levels. Although the original or prevailing concepts associated with individual regional water planes differ somewhat from the views presented here, these differences do not warrant introducing new names. Thus, except for Lake Greens Corners, which is a newly defined level, the lake names used by Chapman (1937) and Stewart (1961) are retained for regional lake features in the study area. On the basis of work at the southern end of the Champlain basin, south of the study area, Connally (1968) suggested the renaming of regional lakes but this problem is beyond the scope of this report.

The terminology for upland lakes in the Winooski and Lamoille Valleys seems hopelessly confused (see literature review by G. G. Connally in this guidebook). For this reason, and because the upland lakes presented here differ substantially in number, extent, elevations, and drainage history from previous reports, new names are used in most cases. Where possible, geographic features near outlet channels associated with newly defined lakes are utilized for the new names. The only exception is Lake Jericho, which was previously named by Connally (1966).

Upland Lakes

Westward recession of the Laurentide ice margin uncovered successively lower outlets, resulting in progressive lowering of lake levels. Lakes Gillett, Huntington, Hollow Brook, Jericho Center, and Jericho developed in that order in the present Winooski drainage basin, and in the present Lamoille basin were Lakes Gillett, Stowe, and The Creek (Figure 2). Lake Gillett is the only lake that extended across the divide between the two present basins. The Lake The Creek outlet channel (T1, Figure 3) extends southward to a delta complex representing Lakes Jericho and Jericho Center (JC2 and J8, Figure 3) indicating general time-equivalence of these lakes. Similarly, the Lake Jericho outlet channel (J1, Figure 3) extends to the Coveville(?) level (C8, Figure 5) in the Champlain Valley, making it possible to relate the upland and regional lake histories.

In addition to the relationship between upland lakes and the Laurentide ice margin, Mountain glacial features can be correlated with the upland lakes, as was previously described (Wagner, 1970). In terms of the lake names used here, Mountain glacier ice margin positions in Ritterbush Valley and North Branch Lamoille River Valley may be contemporaneous with Lake Stowe.

Regional Lakes

The earliest regional lake in the Champlain Valley is represented by the Quaker Springs (?) plane on Figures 1 and 3. The northern extent of this lake probably terminated against the Laurentide ice margin south of Burlington. Slightly older and more southerly ice margin positions in late Quaker Springs (?) time can be inferred by drainage relations. The delta at Bristol (Q1, Figure 1) extends to an outwash surface heading in ice marginal glacial deposits south of Starksboro. The delta near South Hinesburg (Q2, Figure 1) indicates that the Laurentide ice sheet at that time blocked and diverted drainage in the Winooski River Valley through Hollow Brook Valley.

The Coveville (?) water plane (Figure 3) formed immediately after the Quaker Springs level (Stewart, 1961). Chapman's (1937, Figure 16) Coveville plane is shown on Figure 3. The Coveville (?) plane drawn here on Figure 3 is based primarily on features in the Winooski Valley. Although the plane is below Chapman's, it does agree with features identified as Coveville by Connally (1966, 1970) in Vermont and by Denny (1969, personal communication) in New York. The previously described Lake Jericho drainage relations indicate that the Laurentide ice margin blocked the Winooski Valley in Coveville (?) time. Subsequent ice retreat, still in Coveville (?) time, is required for development of Coveville (?) features in the Winooski Valley (Figure 3). Coveville (?) waters may have extended northward to the Lamoille Valley (Connally, 1966), and possibly into Quebec (Parrott and Stone, this guidebook).

The Fort Ann level, first described by Chapman (1937), is the highest regional water-body widely marked by numerous shoreline features on Figure 3. Chapman's (1937, Figures 15 and 16) Fort Ann planes in Vermont and New York, although not coincident, bracket the plane drawn here (Figure 3). The northern extent of the Fort Ann plane is uncertain. According to Chapman (1937, p. 112-113), and Parrott and Stone (this guidebook), the ice margin retreated north of the International Border in late Fort Ann time. McDonald (1968, p. 672-673) tentatively correlated strandline features in the Sherbrooke area of southeastern Quebec with the Fort Ann level. However, if the 230-foot elevation difference between the marine limit and Fort Ann strandlines in the Champlain Valley is compared with data in Quebec, then it appears that McDonald's features are about 25 feet too low to be an extension of the Fort Ann strandline from the Champlain Valley. As discussed below, it may be that Fort Ann time ended when Laurentide ice margin retreat exposed a low divide near Greens Corners, Vermont.

To the south, Fort Ann features, extend beyond the study area (Calkin, 1965; Connally, 1970). Like Chapman's profile, the Fort Ann plane on Figure 3 projects southward to the vicinity of the present Hudson - Champlain divide near Fort Edward, some eight miles south of and at least ten feet higher than Chapman's spillway at Fort Ann, New York.

Below the Fort Ann but above the upper Champlain Sea planes are shoreline features which can be represented by a previously unrecognized water plane (Figure 3). Southward extrapolation of this plane intersects the Champlain Valley floor below the divide, indicating drainage of the lake was northward. To the north the plane extends to a spillway near Greens Corners (Figures 1 and 3). The name "Lake New York" was previously applied (Wagner, 1969) for northward draining lake water immediately below the Fort Ann level and above the Champlain Sea limit, although no specific plane was recognized. Because no evidence for this plane has as yet been found in New York (Denny, 1970, personal communication), the name Greens Corners is applied rather than retain the name Lake New York.

Evidence for a late Pleistocene marine invasion of the St. Lawrence lowland has long been recognized and is generally referred to as the "Champlain Sea"(Karrow, 1961). In the Champlain Valley fossils (chiefly mollusks) and in northern parts "sensitive clay" indicate the presence of saline waters. Chapman (1937, Figure 16) delineated a strandline marking the marine limit, which, as shown on Figure 3, differs somewhat from the fossil-based Champlain Sea maximum of this paper. The only evidence, albeit inconclusive, to support the marine limit based on fossils is the parallelism of this and other water planes, plus close agreement with the marine limit in New York (Denny, 1970, personal communication). A shell date for locality S88(Appendix) basically agrees with the 12,000 year age suggested by McDonald (1968) for the marine maximum.

Below the marine limit Chapman recognized several marine water planes. Although the data on Figure 3 are inconclusive, there are alignments of features approximately coinciding with Chapman's (1937, Figure 16) Port Kent and Burlington levels. In the Winooski Valley deltas are clustered at both the marine limit and at a somewhat lower level (Figure 3) with a pronounced scarp intervening, supporting the Port Kent level (Johnson, 1970). The Port Kent as a level is also supported by shell dates of about 11,300 yrs. B.P. from localities S14 and S24, although there is a discrepancy between shell and wood dates at locality S24 (Appendix). Similarly, age dates from two marine shell localities (nos. S48 and S65) may document the Burlington level as a time line. In Quebec, McDonald (1968, p. 673) found marine shore features were best developed at 115-140 feet below the upper limit, which approximately coincides with Chapman's Port Kent level. However, in northern New York, on the west side of the Champlain Valley, Denny (1969, personal communication) has mapped numerous Champlain Sea features with no apparent stillstand below the marine limit. Recent work with sediments submerged in modern Lake Champlain indicates the end of the Champlain Sea may have occurred about 10,200 years ago (Chase, 1972).

SPECULATIONS

The early work of Chapman established a framework for the late Pleistocene history in northwestern Vermont. This framework is fundamental and likely will stand with little modification. Radiocarbon dates, although only from the Champlain Sea deposits in this area, tend to support Chapman's views. For events preceding and leading up to the Champlain Sea, there is some evidence that the succession of water bodies may not be as straightforward as generally believed. First, some of the deltas at the marine limit in the Missisquoi Valley, and to a lesser extent elsewhere, have complete or nearly complete surface and near-surface veneers of bottom-set sediment (S16; S26; S66; S68; S88). Some other deltas at the marine limit have unusual thicknesses of topset sediment. Secondly, at least two of the marine limit deltas in the Missisquoi Valley have included bodies of till. Thirdly, in the northwestern part of the area are numerous exposures of till overlying a variety of sediments. Figure 4 is a speculative time-space diagram constructed to account for these aspects. As shown, ice recession was accompanied by successive lowering of water levels in the classical fashion, in other words, Quaker Springs, Coveville, Fort Ann, and Champlain Sea. Next, a minor oscillation of the ice margin temporarily reestablished a higher freshwater level, possibly the Fort Ann. At this time some of the previously formed Champlain Sea deltas were submerged and partly veneered with bottom-set sediment and till.

Subsequent recession then lowered the water level to form Lake Greens Corners in the Champlain Valley south of the spillway at

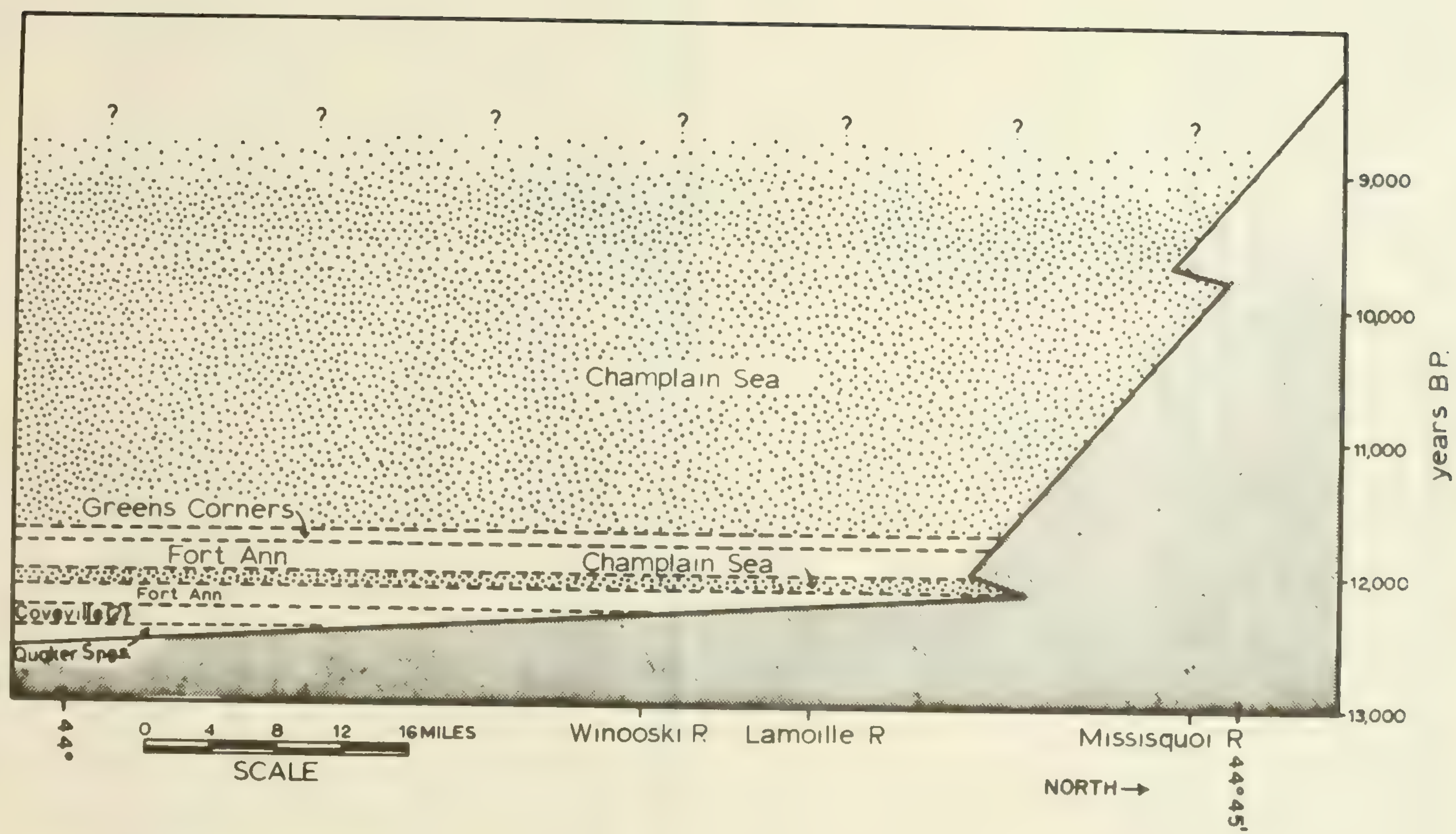

Figure 4: Glacier (shaded) and water body (stippled and plain) relationships in time and space, drawn along 73°10' west longitude meridians.

Greens Corners. Because this spillway extends to a normal delta at the marine limit in the Missisquoi Valley, the Sea by this time apparently had returned to its previous level in that valley. Ice recession finally allowed for the return of the Sea into all of the Champlain Valley, still at its maximum level. Another ice margin oscillation is indicated in later Champlain Sea time, resulting in till deposits in the northwest corner of the state. Figure 5 shows the positions of the ice margins during the times of the positive, southward oscillations. The ice margins are extended into Quebec to show a possible relationship with the Drummondville and Highland Front features (McDonald, 1968).

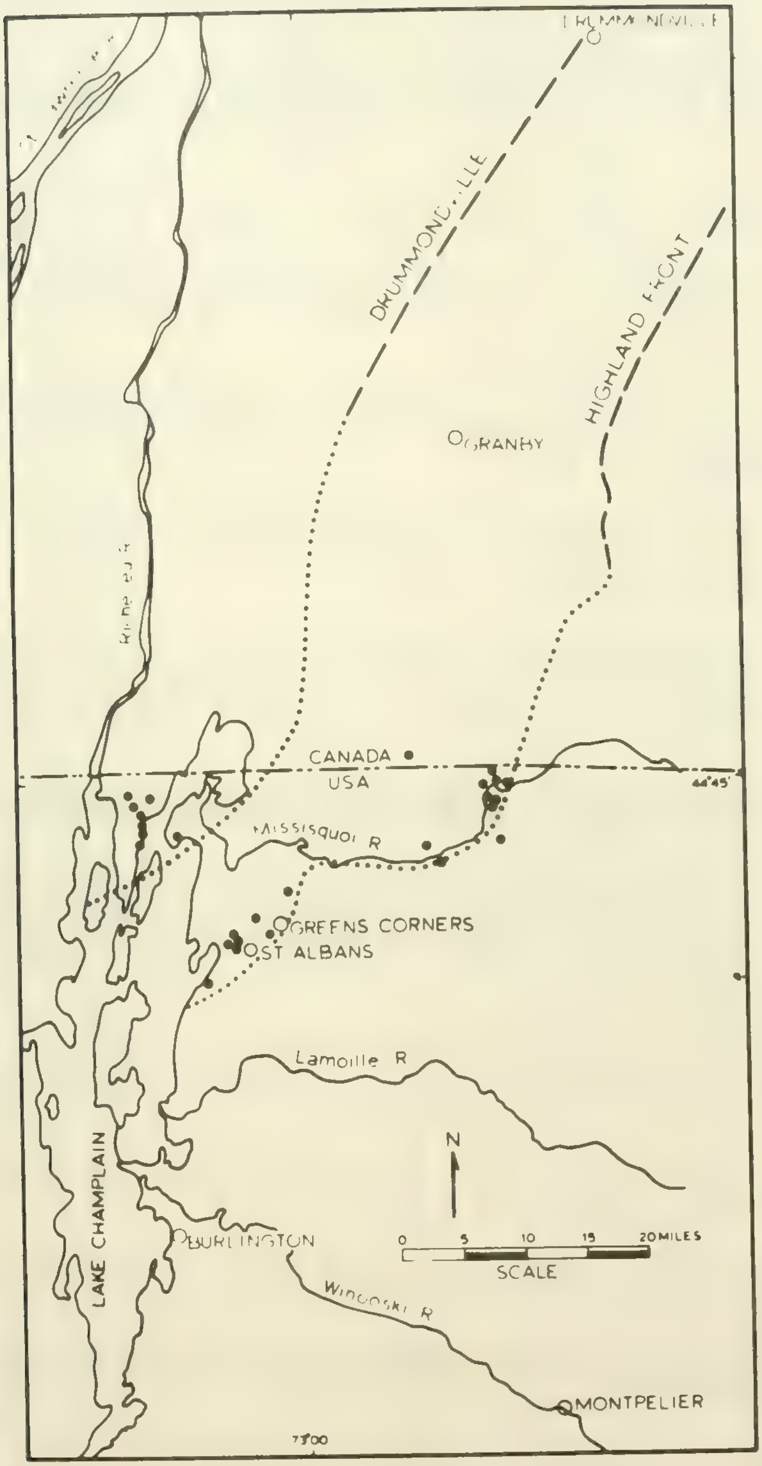

Figure 5; Ice margins in Champlain Valley (dotted lines), and Quebec (dashed lines; from McDonald, 1969, personal communication); Filled circles represent exposures of till overlying non-glacial sediment.

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APPENDIX: Location and Description of Shoreline Features

Feature Name and Number		Type of Feature	Elevation (feet)	Location and Miscellaneous
Gillett	1	outlet channel	760-780	7.5 miles northeast of Gillett Pond; Hunting-ton quad.
	2	delta	800-820	1.5 miles west of Stowe; West Waterbury R.; Mont-pelier quad.
	3	delta	800-820	3.3 miles northeast of Stowe; Glen Bk.; Montpel-ier quad.
	4	delta	820-840	9 miles south of Johnson; Sterling Bk.; Hyde Park quad.
	5	delta	800-820	3.4 miles south of Morris-ville; Bedell Bk.; Hyde Park quad.
Stowe	1	divide	740-760	3.1 miles northeast of Stowe; Montpelier quad.
	2	delta	780-800	Morrisville; Lamoille R.; Hyde Park quad.
	3	delta	780-800	0.8 mile southeast of Johnson; Lamoille R.; Hyde Park quad.
	4	delta	800-840	3 miles northeast of John-son; Gihon R.; Hyde Park quad.
	5	delta	800-820	Belvidere Jct.; North Br., Lamoille R.; Hyde Park quad
The Creek	1	outlet channel	700-720	0.6 mile south of North Underhill; Underhill quad.
	2	delta	720-740	Johnson; Lamoille R.; Hyde Park quad.
	3	delta	700-720	0.7 mile south of Jeffer-sonville; Brewster R.; Jeffersonville quad.
	4	delta	720-740	0.6 mile north of Water-ville; North Br. Lamoille R.; Jeffersonville quad.
	5	delta	740-760	Bakersfield; The Branch; Enosburg Falls quad.
Hunt-ington	1	delta	700-720	Huntington Ctr.; Brush Bk.; Huntington quad.
	2	delta	700-740	1.7 miles southwest of Waterbury; Crossett Br.; Waterbury quad.

Feature Name and Number		Type of Feature	Elevation (feet)	Location and Miscellaneous
Hunt-ington	3	delta	700-720	0.6 mile southeast of Huntington; unnamed stream; Huntington quad.
	4	delta	740-760	1.3 miles northwest of Huntington; unnamed stream; Huntington quad.
	5	delta	720-740	Waterbury Ctr.; Thatcher Bk.; Montpelier quad.
	6	delta	720-740	3.8 miles northwest of Waterbury; Stevenson Br.; Bolton Mtn. quad.
	7	delta	700-760	1.3 miles northwest of Moscow; Miller Bk.; Montpelier quad.
Hollow Brook	1	outlet channel	660-680	3 miles northeast of S. Hinesburg; Hinesburg quad.
	2	delta	660-680	4.5 miles northeast of S. Hinesburg; unnamed stream; Hinesburg quad.
	3	delta	680-700	1.3 miles south of Waterbury Ctr.; Thatcher Br.; Montpelier quad.
	4	delta	700-720	3.8 miles northwest of Waterbury; Stevenson Br.; Bolton Mtn. quad.
Jericho	1	delta	620-640	Huntington; Huntington R.; Huntington quad.
	2	delta	640-660	Waterbury; Winooski R.; Montpelier quad.
	3	delta	620-640	1.2 miles northwest of Huntington; unnamed stream; Huntington quad.
	4	delta	620-640	1.5 miles northwest of Huntington; unnamed stream; Huntington quad.
	5	delta	620-640	2.8 miles northwest of Huntington; unnamed stream; Huntington quad.
	6	delta	640-660	3.8 miles northwest of Waterbury; Stevenson Bk.; Bolton Mtn. quad.
	7	outlet channel	660-680	1.9 miles southwest of Williston; Essex Jct.quad.
	8	delta	690	1 mile northeast of Jericho; Browns R.; Underhill quad.

Feature Name and Number		Type of Feature	Elevation (feet)	Location and Miscellaneous
Jericho Center	1	outlet channel	680-700	Jericho Center; Richmond quad.
	2	delta	706	Underhill; Browns R. and The Creek; Underhill quad.
Quaker Springs (?)	1	delta	560-580	Bristol; New Haven R.; Bristol quad.
	2	delta	600-620	0.4 mile southeast of S. Hinesburg; Hollow Brook; Hinesburg quad.
Coveville (?)	1	delta	540-580	1.5 miles northwest of Richmond; Winooski R.; Essex Jct. quad.
	2	delta	580-600	1.5 miles north of Richmond; Mill Bk.; Richmond quad.
	3	delta	560-600	0.4 mile south of Williston; Allen Bk.; Essex Jct. quad.
	4	delta	580-600	2.6 miles northwest of Richmond; Winooski R.; Essex Jct. quad.
	5	delta	606	0.9 mile southwest of Jericho Center; unnamed brk.; Richmond quad.
	6	delta	600-620	1.6 miles southeast of Jericho; Lee R.; Richmond quad.
	7	delta	620-640	at Jericho; Browns R.; Underhill quad.
	8	delta	600-620	
Fort Ann	1	delta(?)	380-400	0.9 mile east of New Haven Mills; unnamed stream; South Mtn. quad.
	2	beach	390-400	2.3 miles southeast of Vergennes; west side of Buck Mtn.; Monkton quad.
	3	delta	400-420	0.6 mile south of Bristol; New Haven R.; Bristol quad.
	4	spit	400-410	4.1 miles northwest of Bristol; Monkton quad.
	5	delta	420-480	0.8 mile east and northeast of Hogback Mtn.; Hinesburg quad.
	6	beach	400-420	1.5 miles east of N. Ferrisburg; Mt. Philo quad.
	7	beach	440-460	southwest side of Mt. Philo; Mt. Philo quad.

Feature Name and Number		Type of Feature	Elevation (feet)	Location and Miscellaneous
Fort Ann	8	beach	420-440	southwest side of Mt. Philo; Mt. Philo quad.
	9	beach	400-420	southwest side of Mt. Philo; Mt. Philo quad.
	10	beach	360-380	southwest side of Mt. Philo; Mt. Philo quad.
	11	delta	380-400	1.9 miles southwest of S. Hinesburg; Lewis Creek; Hinesburg quad.
	12	delta	460-500	South Hinesburg; Hollow Brook; Hinesburg quad.
	13	beach	400-460	south side of Pease Mountain; Mt. Philo quad.
	14	delta	360-380	1.9 miles southeast of Hinesburg; LaPlatte R.; Hinesburg quad.
	15	beach	480-500	four unnamed hillocks about 1.3 miles east of E. Charlotte; Mt. Philo quad.
	16	beach	440-460	south side of Jones Hill; Mt. Philo quad.
	17	beach	440-500	0.8 mile northeast of East Charlotte; Mt. Philo quad.
	18	bench	480-510	0.2 mile north of Rts. 116 and 2A, intersection and north along Rt. 116; Mt. Philo and Burlington quads.
	19	delta	500-520	Williston; Winooski R.; Essex Jct. quad.
	20	delta	480-520	1.1 miles east of Essex Jct.; Winooski River; Essex Jct. quad.
	21	delta	500-540	0.2 mile south of Jericho Cemetery; Lee R.; Underhill quad.
	22	delta	500-525	Essex Center; Alder Brook; Essex Center quad.
	23	delta	530-550	Brookside Cemetery; Rogers Brook; Essex Center quad.
	24	beach	520-540	southeast side of Cobble Hill; Fort Ethan Allen quad.
	25	beach	520-580	1.3 miles west of Milton Pond; Milton quad.
	26	delta	520-540	2.5 miles north of Westford; Browns River; Gilson Mtn. quad.

Feature Name and Number		Type of Feature	Elevation (feet)	Location and Miscellaneous
Fort Ann	27	delta	540-580	Fairfax Falls; Lamoille R.; Gilson Mtn. quad.
	28	delta	540-560	River View School; Lamoille R.; Gilson Mtn. quad.
	29	beach	520-560	east side of Arrowhead Mtn.; Milton quad.
	30	delta	560-580	Binghamville; Stones Brook; Gilson Mtn. quad.
	31	delta	600-620	Buck Hollow; esker-fed; Milton quad.
	32	beach	590-610	0.7 mile southwest of Bellevue Hill; St. Albans quad.
Greens Corners	1	delta	200-220	Weybridge; Otter Creek; Middlebury quad.
	2	beach	240-250	0.8 mile southeast of Vergennes; Monkton quad.
	3	beach	230-250	0.8 mile northeast of Vergennes; Monkton quad.
	4	beach	260	0.8 mile northeast of Ferrisburg; Monkton quad.
	5	delta	280-300	0.5 mile southwest of North Ferrisburg; Lewis Creek; Mt. Philo quad.
	6	beach	260-280	0.1 mile northwest of Coleman Corner; Mt. Philo quad.
	7	beach	300-320	0.2 mile north of Coleman Corner; Mt. Philo quad.
	8	beach	280-300	0.9 mile west of Mt. Philo; Mt. Philo quad.
	9	delta(?)	300	1 mile south of Prindle Corners; Lewis Creek; Mt. Philo quad.
	10	beach	280-300	0.3 mile southeast of Barber Hill; Willsboro quad.
	11	delta(?)	320-340	0.4 mile northwest of Hinesburg; LaPlatte R.; Hinesburg quad.
	12	beach	380-400	1.9 miles southeast of Essex Jct.; Essex Jct. quad.
	13	delta	360-380	1.4 miles southeast of Essex Jct.; Winooski R.
	14	delta	420-440	Fairfax; Lamoille R.; Milton quad.
	15	delta & outlet channel	500-510	1.5 miles northeast of Greens Corners; St. Albans quad.

Feature Name and Number	Type of Feature	Elevation (feet)	Location and Miscellaneous
Champlain Sea 1	delta	100	1.5 miles south of West Bridport; Crown Pt. quad. mollusks; 9,620 ± 350 B.P. shell date I-4695.
2	delta	175	1.3 miles southwest of Weybridge; Middlebury quad.
3	beach	180-200	3 miles north of Addison; Port Henry quad.
4	beach	200-210	≈.7 mile northwest of Buck Mt.; Monkton quad.
5	delta	160-180	1.6 miles west of Vergennes; Port Henry quad.
6	delta	120-140	2 miles northeast of Panton; Port Henry quad.
7	beach	200-210	.2 mile northeast of Ferrisburg; Monkton quad.
8	beach	200-210	1.9 miles northeast of Ferrisburg; Monkton quad.
9	delta	100-120	≈1 mile east of Hawkins Bay; Port Henry quad.
10	delta	200-220	1.2 miles southwest of North Ferrisburg; Mt. Philo quad.
11	beach	200-220	1.9 miles northwest of North Ferrisburg; Mt. Philo quad.
12	delta	160-180	1.5 miles west of North Ferrisburg; Mt. Philo quad.
13	beach	160-180	2.5 miles south of Charlotte; Willsboro quad.; mollusks.
14	beach	180-200	1.8 miles southeast of Charlotte and west of Thompsons Point; Willsboro quad.; mollusks; 11,230±170 B.P. shell date I-3647.
15	beach	240-260	.6 mile southwest of Jones Hill cemetery; Mt. Philo quad.
16	delta	260-300	1.9 miles southeast of Shelburne Falls; Mt. Philo quad.
17	delta	260-280	.9 mile south of Shelburne Falls; Mt. Philo quad.

Feature Name and Number	Type of Feature	Elevation (feet)	Location and Miscellaneous
Champlain Sea 18	delta	200-220	.3 mile west of Shelburne Falls; Mt. Philo quad.
19	beach	200-220	1.8 miles east of Shelburne; Burlington quad.; mollusks.
20	delta	140-160	.3 mile northeast of Shelburne; Burlington quad.
21	delta	100-120	1.5 miles northwest of Shelburne; Burlington quad.
22	beach	140-300	.7 mile southeast of Twin Orchards; Burlington quad.
23	beach	200-270	.5 mile southeast of Queen City Park; Burlington quad.
24	beach	180-200	1.8 miles northeast of Queen City Park; Burlington quad.; mollusks and wood; 10,950$\pm$300 B.P. wood date W-2309; 11,420 $\pm$350 shell date W-2311.
25	beach	280-300	1.5 miles southwest of South Burlington; Burlington quad.
26	delta	280-300	2 miles southeast of South Burlington on Rte. 2; Burlington quad.
27	delta	320-340	.3 mile west of Ft. Ethan Allen Military Res.; Ft. Ethan Allen quad.
28	delta	320-340	1.1 miles northwest of Ft. Ethan Allen; Ethan Allen quad.
29	delta	300-320	1.5 miles northwest of Ft. Ethan Allen; Ethan Allen quad.
30	delta	180-200	.4 mile east of Shipman Hill; Ft. Ethan Allen quad.
31	delta	160-180	.4 mile southwest of Bayside; Ft. Ethan Allen quad.
32	beach	320-340	1.5 miles southwest of Colchester; Ft. Ethan Allen quad.

Feature Name and Number	Type of Feature	Elevation (feet)	Location and Miscellaneous
Champlain Sea 33	delta	300-320	1.4 miles east of Colchester; Ft. Ethan Allen quad.
34	beach	170-190	1.2 miles west of Bayside; Ft. Ethan Allen quad; mollusks.
35	delta	120-140	1.2 miles from tip of Malletts Head; Ft. Ethan Allen quad.
36	beach	200-220	.8 mile from tip of Malletts Head; Ft. Ethan Allen quad.
37	delta	300-320	1.4 miles north of Colchester Pond; Essex Center quad.
38	delta	190-200	.8 mile northwest of Chimney Corner; Ft. Ethan Allen quad.
39	beach	250-270	.9 mile northwest of Walnut Ledge; Ft. Ethan Allen quad. mollusks.
40	delta	320-340	at Checkerberry Village; Georgia Plains; mollusks; 10,520±180 B.P. shell date I-4393.
41	delta	360-380	.8 mile south of Arrowhead Mtn.; Milton quad.
42	delta	300-320	.7 mile south of Towns Corner School; Georgia Plains quad.
43	beach	180-200	.6 mile southwest of Silvertown School; Georgia Plains quad.
44	delta	380-400	.4 mile north of Arrowhead Mountain Lake; Milton quad.
45	delta	200-220	.7 mile east of Miltonboro; Georgia Plains quad.
46	delta	180-200	.1 mile north of Miltonboro; Georgia Plains quad.
47	beach	170-190	.6 mile northwest of Miltonboro; Georgia Plains quad.; mollusks.
48	beach	160-200	1.2 miles northwest of Miltonboro; Georgia Plains quad.; mollusks; 10,460±180 B.P.; shell date I-4394.

Feature Name and Number	Type of Feature	Elevation (feet)	Location and Miscellaneous
Champlain Sea 49	beach	300-320	2.5 miles southeast of Georgia Plains; Georgia Plains quad.; mollusks.
50	beach	190-220	1.5 miles west of Georgia Plains; Georgia Plains quad.; mollusks.
51	delta	240-260	at Georgia Plains; Georgia Plains quad.
52	delta	230-240	.6 mile southeast of Melville Landing; St. Albans Bay quad.
53	delta	180-200	1 mile northeast of Lime Rock Pt.; St. Albans Bay quad.
54	delta	380-400	at East Fairfield; Enosburg Falls quad.
55	beach	380-400	.6 mile west of Holy Cross Cemetery; St. Albans quad.
56	delta	380-400	2.5 miles northwest of East Fairfield; Enosburg Falls quad.
57	beach	300-320	1 mile west of Holy Cross Cemetery; St. Albans quad.
58	beach	180-200	2 miles northwest of St. Albans; St. Albans quad.
59	beach	390-400	.1 mile east of WWSR radio tower; St. Albans quad.
60	delta	380-400	.5 mile north of Fairfield Station; Enosburg Falls quad.
61	beach	310-320	1.5 miles south of Fonda; St. Albans quad.
62	beach	220-230	.7 mile south of Fonda; St. Albans quad.
63	beach	180-200	at gravel pit Morin Road south of Swanton; East Alburg quad.
64	beach(?)	100-120	1.5 miles southeast of Town of Isle La Motte; Rouses Point quad.;mollusks.
65	beach(?)	180-200	.7 mile north of Town of Isle La Motte; Rouses Point quad.; mollusks.
66	delta	400-420	at Sheldon; Enosburg Falls quad.
67	delta	420-440	at Enosburg Falls; Enosburg Falls quad.

Feature Name and Number	Type of Feature	Elevation (feet)	Location and Miscellaneous
Champlain Sea 68	delta	420-440	.5 mile south of Enosburg Falls; Enosburg Falls quad.
69	delta	380-400	at South Franklin; Enosburg Falls quad.
70	delta	300-320	1 mile west of Sheldon Springs; Enosburg Falls quad.
71	delta	440-460	Enosburg Falls; Enosburg Falls quad.
72	delta	440-460	at East Berkshire; Jay Peak quad.
73	delta	300-310	1.1 miles east of Highgate Ctr.; Highgate Ctr. quad.
74	delta	230-250	1.5 miles east of Swanton; Highgate Ctr. quad.
75	beach	200-210	.9 mile east of Swanton; Highgate Ctr. quad.
76	beach	189	.6 mile east of Swanton; Highgate Ctr. quad.
77	beach	160-180	1.5 miles west of Bluff Point; Rouses Point quad.
78	delta	150-160	at Swanton; Highgate Ctr. quad.
79	beach	120-130	1.3 miles west of Swanton; East Alburg quad.
80	delta	120-140	.4 mile north of Swanton; East Alburg quad.
81	beach	140	1.1 miles north of Swanton; Highgate Ctr. quad.
82	delta	100-120	1.4 miles northwest of Swanton; East Alburg quad.
83	delta	100-120	.5 mile west of Blue Rock; Rouses Point quad.;mollusks(?)
84	beach	120-130	1.2 miles northeast of West Swanton, East Alburg quad.; mollusks.
85	beach	300	1.3 miles southwest of Center Pond; Highgate Ctr. quad.; mollusks.
86	delta	460-480	.9 mile southwest of Richford; Jay Peak quad.
87	delta	440-460	.25 mile north of North Enosburg; Enosburg Falls quad.
88	delta	475	2 miles south of Frelighsburg, Quebec; mollusks; 11,740$\pm$200 B.P. shell date I-4489.

Feature Name and Number	Type of Feature	Elevation (feet)	Location and Miscellaneous
Miscellaneous			
1	spit?	500-520	Mount Philo; Mt. Philo quad.
2	kame-delta	700-720	1.1 miles east of South Hinesburg; Hinesburg quad.
3	delta	640-660	.7 mile east of South Hinesburg; Hinesburg quad.
4	kame-delta	880-900	1.3 miles northeast of Jonesville; Richmond quad.
5	kame-delta	740-760	1 mile west of Oak Hill School; Essex Jct. quad.
6	kame-delta	720-740	1.1 miles south of Jericho Ctr.; Richmond quad.
7	kame-delta	900-920	2.3 miles east of Lake Mansfield; Bolton quad.

PROGLACIAL LAKES IN THE LAMOILLE VALLEY, VERMONT

by

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Three proglacial lakes were present in the Lamoille Valley during, and following, retreat of the late Woodfordian glacier in the Champlain Valley. This glacier deposited the Burlington drift of Stewart and MacClintock (1969). Although these lake levels have been recognized since the early part of this century, the nomenclature is still confused, as seen in Table 1. This discussion is a summary of previously published works of others, and of field work performed sporadically for the past six years. Because the names Lake Lamoille and Lake Mansfield have priority in the Lamoille Valley, they are retained in this paper.

TABLE 1.

MERWIN, 1908	CHAPMAN, 1937 1942	STEWART, 1961
Lake Lamoille I	---	Lake Mansfield
Lake Mansfield	---	Lake Lamoille
Lake Lamoille III	Coveville Stage (Lake Vermont)	Coveville Stage (Lake Vermont)

CONNALLY, 1966 1968	STEWART AND MACCLINTOCK, 1969	CONNALLY, 1971
Lake Lamoille	Quaker Springs Stage ? (Lake Vermont)	Lake Lamoille
Lake Mansfield	Quaker Springs Stage ? (Lake Vermont)	Lake Mansfield
Coveville Stage (Lake Vermont)	Coveville Stage (Lake Vermont)	Lake Coveville

Merwin (1908) recognized an upper level (above 800'), designated Lake Lamoille I, that he thought had been restricted to the Lamoille Valley. He proposed that the lowland east of Mount Mansfield, between Morrisville and Stowe, was then cut down by steadily lowering lake waters, designated Lake Lamoille II. Then, when the outlet was breached to its present level (740') the waters of Lake Lamoille II and Lake Winooski I, in the Winooski Valley to the south, joined to form Lake Mansfield. The lowest level in the Lamoille Valley (650'), presumed to have been restricted to that valley, was named Lake Lamoille III. Fairchild (1916) recognized Merwin's terminology in the Lamoille Valley except that he erroneously projected his upper marine limit (the Champlain Sea) in place of Lake Lamoille III. Chapman (1937, 1942) projected the Coveville Stage of Lake Vermont to Merwin's Lake Lamoille II features, an interpretation that has been generally recognized to the present, the only change being the redesignation as Glacial Lake Coveville by Connally and Sirkin (1970). In mapping the bedrock geology of the Mount Mansfield quadrangle Christman (1959, p. 73) clearly recognized the priority of Merwin's terms although he chose "Lake Lamoille deposits"(quotations his) as a mapping unit.

Stewart (1961) correctly inferred that the upper lake actually extended into the Winooski Valley and was not restricted to the Lamoille Valley as Merwin (1908, p. 132) had supposed. He also inferred that the lower lake did not - an interpretation supported here - also contrary to the concepts of Merwin (ibid, p. 136). Stewart therefore honored the conceptual priority and renamed the upper lake, Lake Mansfield, and the lower, Lake Lamoille, reversing Merwin's terms. Connally (1966, 1968), however, re-established Merwin's names, concluding that the original elevations and features were the most important precedent. Then, Stewart and MacClintock (1969) thoroughly confused matters by reapplying the names Lake Lamoille and Lake Mansfield to problematical higher levels and by apparently assigning both of Merwin's levels to the Quaker Springs Stage of Lake Vermont, even though these lakes are not at the proper elevations (Connally, 1966, 1968, and elsewhere) for the Champlain Valley lake.

Merwin's original terminology is retained and defended here for three reasons: (1) these terms were accepted for more than 50 years prior to the work of Stewart, (2) these terms were applied to specific features and elevations that have been studied and restudied for more than 60 years, and (3) it is less confusing to either extend (Lake Lamoille) or restrict (Lake Mansfield) existing terms, when they are meaningful, than to introduce new names because of original conceptual flaws.

GLACIAL LAKE LAMOILLE

This lake is defined by six deltas in the Lamoille Valley and at least four between Morrisville and Stowe, east of Mount

Mansfield. Two of the deltas near Stowe were originally mapped by Wagner (1970, personal communication). The Lake Lamoille deltas (Figure 1) range from 840' in the northwest to 780' in the southeast, as determined from flat delta tops depicted on 7 1/2' topographic maps. Lake Lamoille was blocked by the ice margin in the west and drained southward via the Winooski Valley. Wagner has located the outlet for this lake at about 760' at Gillett at the west end of the Winooski Valley. Figure 2 shows a projection of Lakes Lamoille, Mansfield, and Coveville along A-A' in Figure 1.

GLACIAL LAKE MANSFIELD

This lake is defined by seven deltas and two beaches. The deltas (Figure 3) range from 760' in the north to 720' in the south. Merwin suggested that this lake coalesced with one in the Winooski Valley, however, the divide may be about 20' too high to have permitted this (Figure 2). I suggest that initial drainage was through the Stowe lowland, while the ice blocked the valley of The Creek west of Mount Mansfield. Later, the ice block was dissected in The Creek and this channel controlled falling lake levels. The The Creek channel is at 700' and no shoreline features are graded to this elevation so it must have controlled a very short-lived lake level. Since Lake Mansfield is now defined only in the Lamoille Valley, this restricts the original definition of Merwin (1908).

GLACIAL LAKE COVEVILLE

This lake is documented by nine deltas and two beaches (Figure 4) that range from 660' to 640' at Morrisville. The inclusion of these features with Lake Coveville has never been challenged but it is fraught with problems as discussed by Wagner (1969). Connally and Calkin (1972) document the retreat of an active ice margin during Lake Coveville, including the Bridport readvance that took place between Burlington and Bridport(south of Middlebury). The retreating margin of an active glacier may account for many of the problems outlined by Wagner. A projection of Lamoille Valley features onto a generalized north-south Lake Coveville projection in the Champlain Valley strongly supports coincidence of the levels (Figure 5).

TIME STRATIGRAPHY

In Figure 5 a hypothetical projection of Lake Quaker Springs is shown. Both Lake Lamoille and Lake Mansfield had to drain southward into the Champlain Valley. If the projections are correct, Lake Mansfield must have drained into Lake Coveville (via Lake Jericho in the Winooski Valley) and not Lake Quaker Springs. Perhaps

Lake Mansfield was dammed by the Bridport readvance after a period of free drainage. Differential rebound (Figure 2) between Lake Lamoille and Lake Mansfield suggests that some event separated the two lakes and that Lake Lamoille drained through a series of impondments into Lake Quaker Springs at its northern boundary near Brandon.

Connally and Sirkin (1972) have estimated the age of Lake Coveville as 12,800 yrs. B.P. and the Luzerne readvance, that they tentatively correlated with the Burlington drift, as 13,200 yrs. B.P. Thus, it is probable that Lakes Lamoille and Mansfield existed sometime between 13,200 and 12,800 yrs. B.P. Because two of the local mountain glaciers reported by Wagner (1970) can be directly related to Lake Lamoille; one in the Ritterbush Valley and one east of Belvidere Center, it is probable that these glaciers also existed between 13,200 and 12,800 yrs. B.P.

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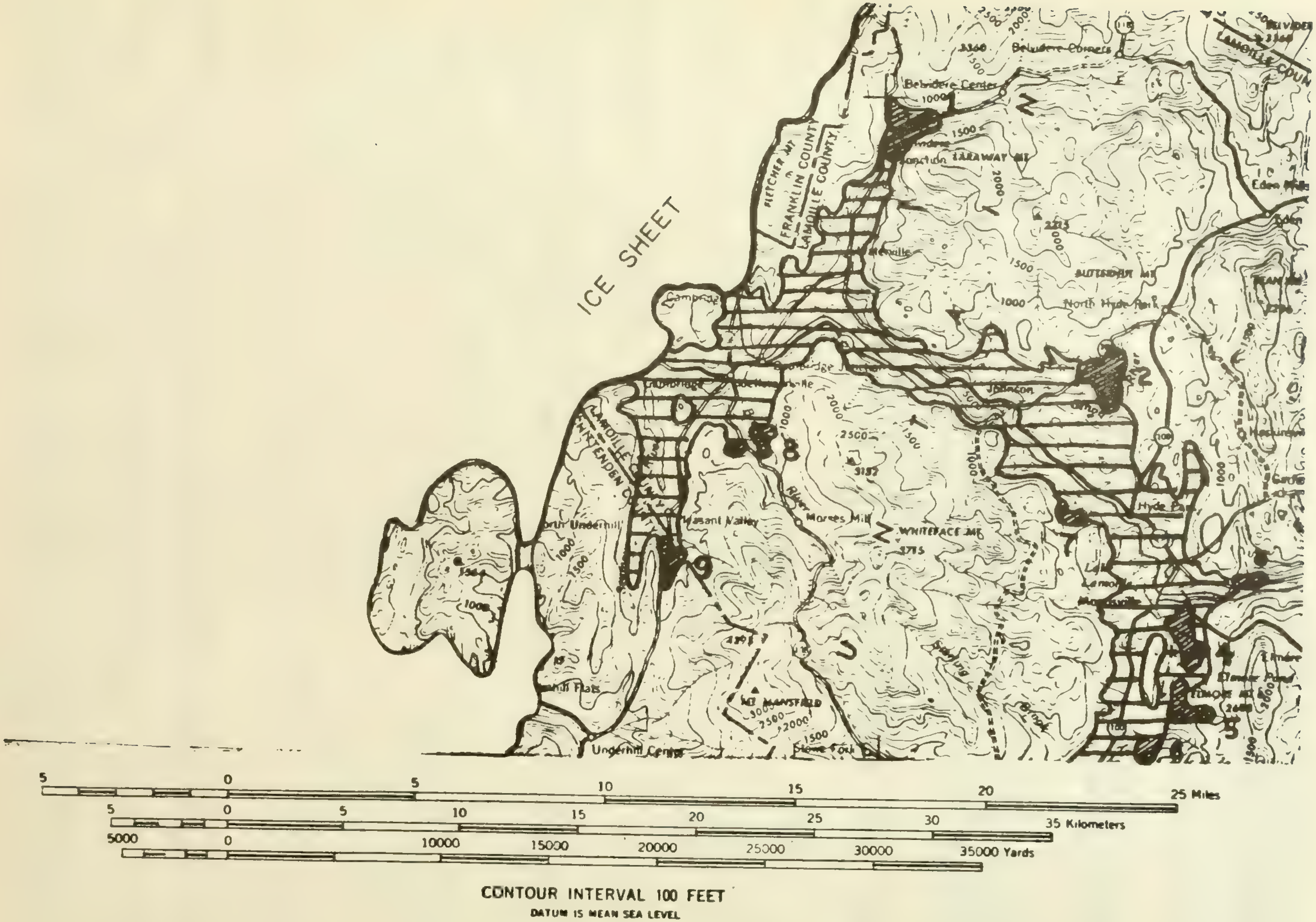

Figure 1. Glacial Lake Lamoille, showing numbered deltas and the probable extent of continental ice that dammed this lake to the west. Numbers correspond to those in Figure 4.

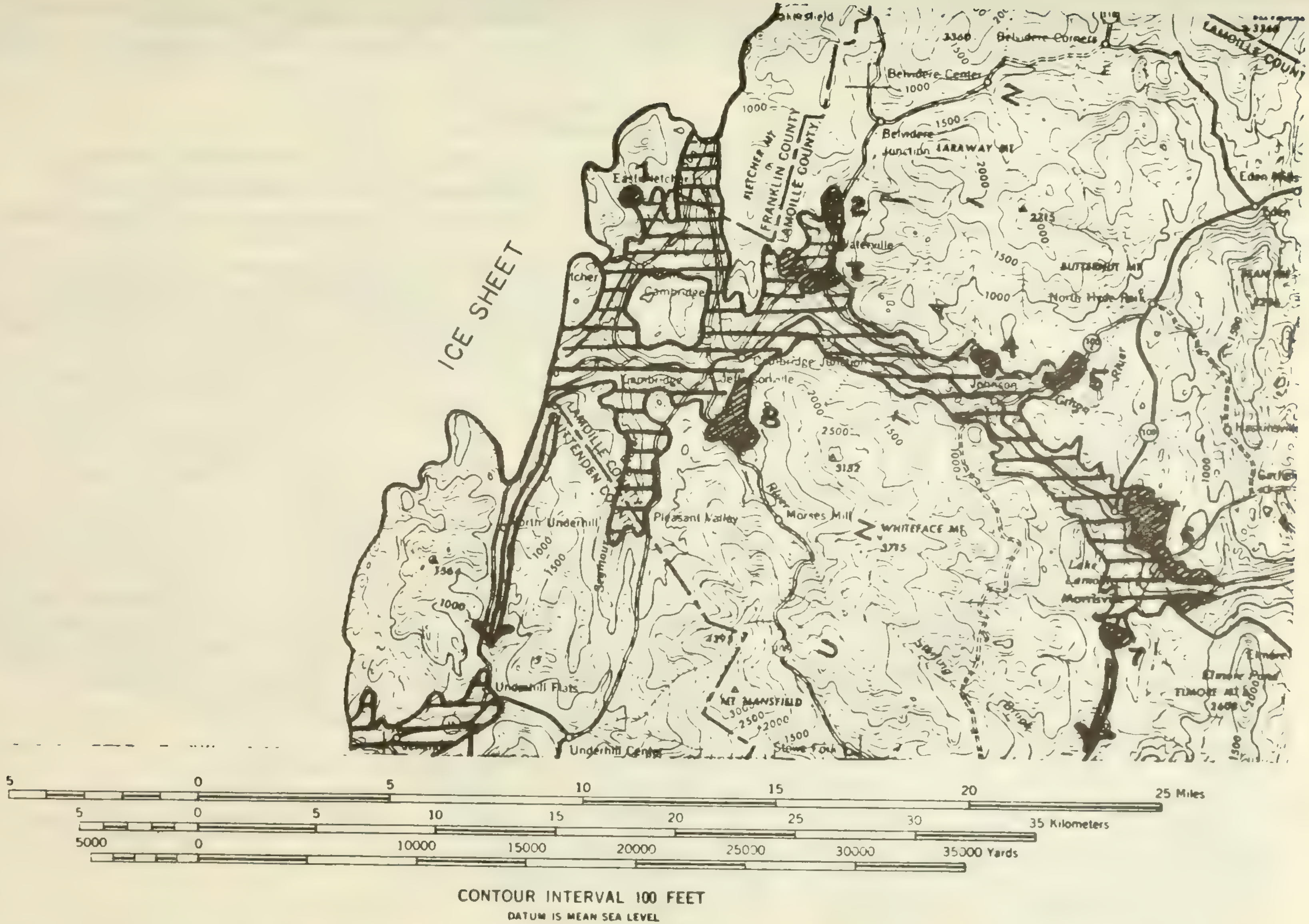

Figure 2. Glacial Lake Mansfield, showing numbered deltas and beach features and the probable extent of continental ice that dammed this lake to the west. Glacial Lake Jericho is also shown southwest of Lake Mansfield and probable drainage lines through The Creek in the west and over the Stowe outlet in the east are depicted by arrows. Numbers correspond to those in Figure 4.

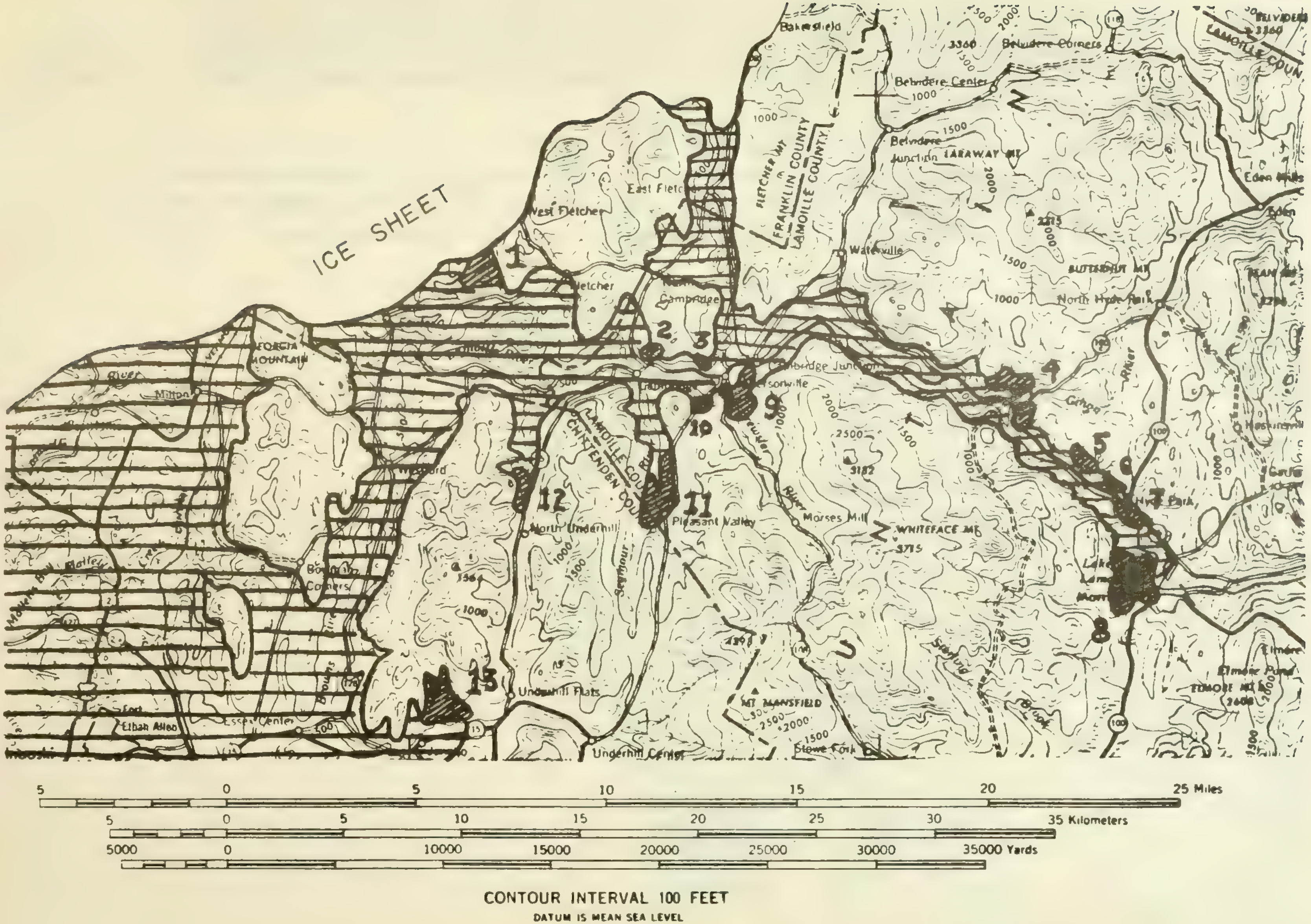

Figure 3. Glacial Lake Coveville, an early phase, showing numbered deltas and beach features. Numbers correspond to those in Figure 4.

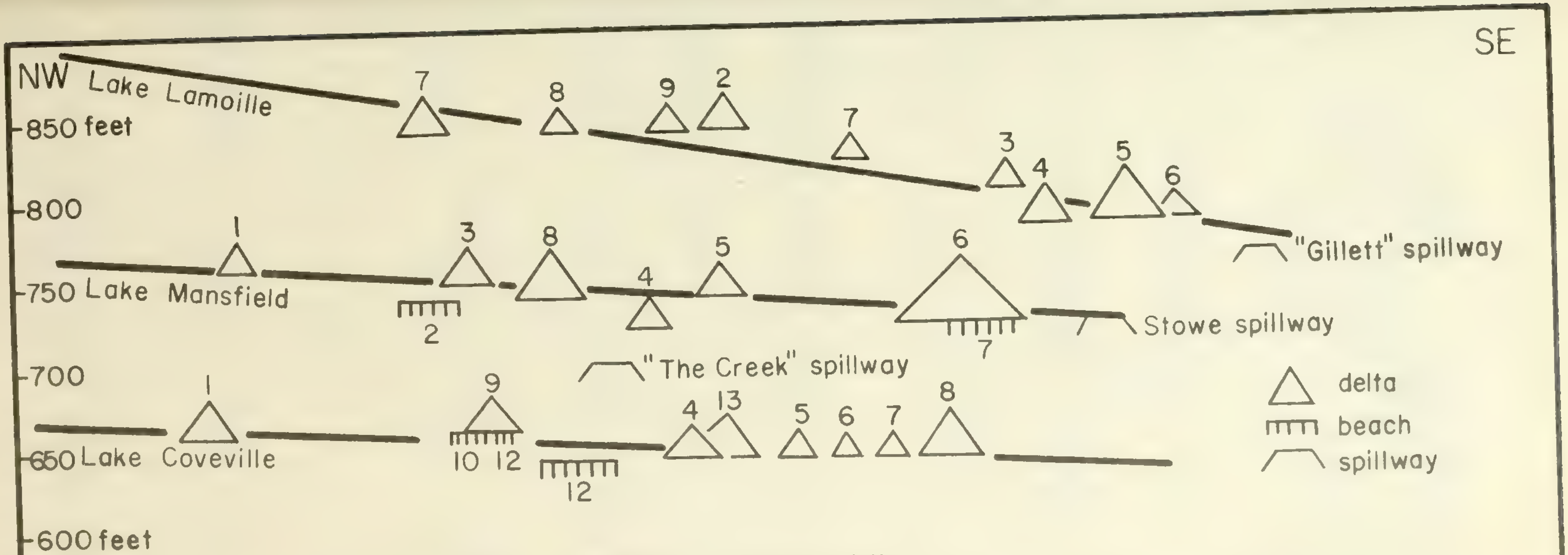

Figure 4. Projected Lake Levels, Lamoille Valley

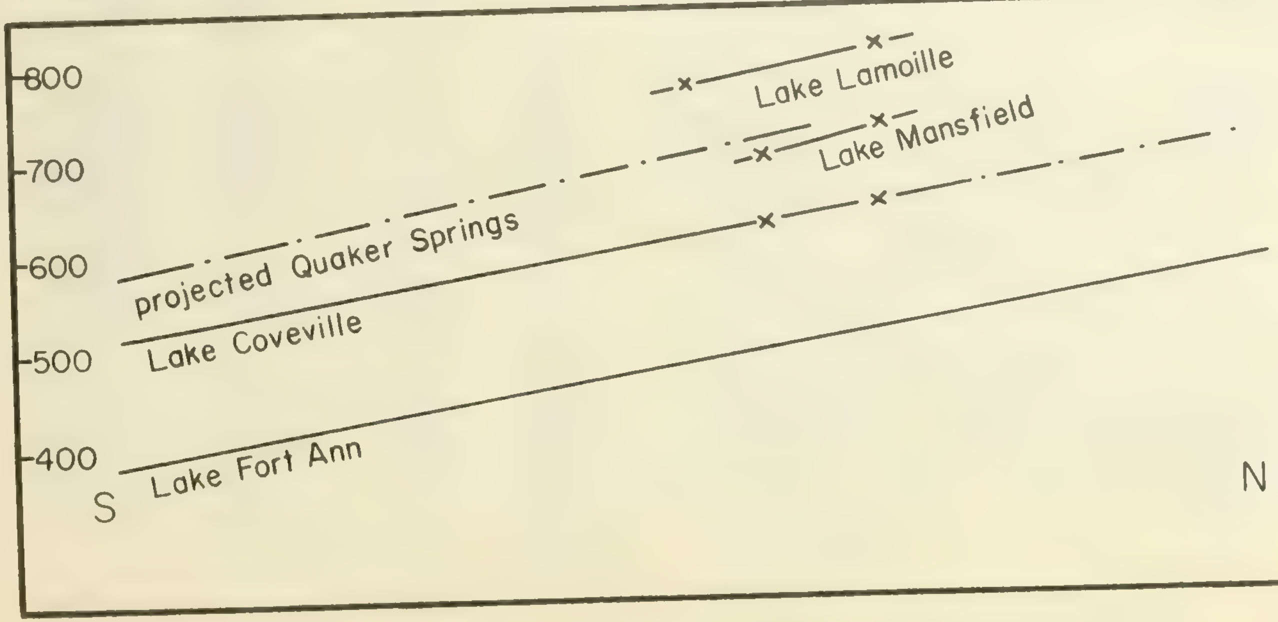

Figure 5.

North-South projections for the Champlain Valley; X's are projected elevations from Lamoille Valley lakes.

CONNALLY - WAGNER ROAD LOG

Miles between Points	Cumulative Mileage	Description
0.0	0.0	START Spear St. and Route 2.
0.3	0.3	I-89
1.2	1.5	Crossing the Winooski River.
0.6	2.1	Winooski Exit.
0.6	2.7	St. Michaels College - Champlain Sea delta at the marine limit.
1.2	3.9	Fort Ethan Allen (retired) on Champlain Sea delta.
1.0	4.9	Essex Junction on Champlain Sea delta.
0.3	5.2	North side of Essex Junction: note gully on contact between delta and till and lake sediment veneered bedrock upland.
2.7	7.9	Essex Center on Lake Fort Ann delta.
1.7	9.6	Descend from Lake Fort Ann delta to Browns River terrace.
0.4	10.0	Cross Browns River.
0.4	10.4	Lake Fort Ann delta remnant on left on till covered upland; note residual boulders in gully.
0.4	10.8	Lake Fort Ann surface at right across Browns River.
0.3	11.1	Cross Browns River.
0.2	11.3	Village of Jericho on Lake Coveville delta.
0.5	11.8	Ascend Lake Jericho delta.
1.7	13.5	Leave Lake Jericho delta and continue on the south terrace of Browns River that is graded to Lake Jericho.

Miles between Points	Cumulative Mileage	Description
0.2	13.7	Cross Browns River and ascend matching terrace on north; village of Underhill. Two sequences of ice contact drift on hillside on right.
1.5	15.2	Cross The Creek; kame terraces on both right and left valley walls.
1.7	16.9	Ice contact drift (kame terraces) on left and divide between south-flowing The Creek and a north-flowing tributary of the Lamoille River on the right. This divide is crucial in the correct interpretation of Lamoille Valley lakes. The elevation is approximately 700 ft.; too low for Lake Mansfield (720-740 ft.); and too high for Lake Coveville (640-660 ft.). Wagner has proposed this as an outlet for a lake he has named Glacial Lake The Creek. Clearly continental ice blocked this col during Lake Lamoille (840 ft.) and at least initial Lake Mansfield, and either retreated or was breached prior to the establishment of Lake Coveville in the Champlain Valley.
1.8	18.7	North Underhill; head of proposed spillway for Lake The Creek. Ice contact drift in valley bottom and on valley walls.
5.8	24.5	Village of Cambridge; village is very close to 10 year floodplain.
0.6	25.1	Cross Lamoille River.
1.9	27.0	Cross Lamoille River; village of Jeffersonville. Follow Route 108 south.
1.6	28.6	STOP 1 (Connally): Three lake levels can be inferred from the stream terraces and delta remnants in the Brewster River valleys.

Miles between Points	Cumulative Mileage	Description
		The lowest surface, to the north, has a sharp slope break at 660 ft. The one on which we stand has a break at 740 ft. Higher terraces are graded to 840 ft. and a small delta remnant may be present at that elevation. The upper level has been assigned to Lake Lamoille, the intermediate to Lake Mansfield, and the lowest to Lake Coveville. Here we will discuss the possible relationship between these lake levels and the The Creek divide.
0.8	29.4	A 20 ft. high erosional scarp in the terraces graded to the 740 ft. delta.
0.8	30.2	Village of South Cambridge; ascend the terrace graded to the 840 ft. level.
1.9	32.1	Gravel pits that showed forset beds in 1965 and bottomset beds in 1970. This delta documents an early local lake at about 1100 ft. dammed by the retreating continental ice margin.
2.6	34.7	Protalus rampart(?) at north entrance to Smugglers Notch; abundant talus and mudslide debris.
2.9	37.6	Stream exposures of ice contact drift and till; collapse structures.
1.0	38.6	Kame deltas(?) or kame moraine(?) in vicinity of Toll House Inn, headwaters of the Waterbury River.
3.8	42.4	Holme Lodge - valley bottom floored with more than 100 ft. of unconsolidated material.
0.2	42.6	Leave Route 108; make sharp right turn and follow signs to Trapp Family Lodge.

Miles between Points	Cumulative Mileage	Description
0.5	43.1	Ten Acres Lodge on 800 ft. delta assigned to Glacial Lake Gillett by Wagner.
1.6	44.7	STOP 2 (Wagner): Trapp Family Lodge. Just beyond Lodge is good view of Miller Brook Valley. Photo stop.
1.7	46.4	Continue on dirt road to black top, make right turn immediately, onto dirt surface. Cross Miller Brook and take first right.
1.8	48.2	STOP 3 (Wagner): Phase I Mountain glaciation. Park cars in field across from house and walk up dirt road onto delta surface. Delta was constructed from outwash with stagnant ice margin up valley. Proceed up valley to Lake Mansfield Trout Club.
2.2	50.4	STOP 4 (Wagner): Phase II Mountain glaciation. Walk across dam breast and follow white blazed trail to lateral moraine. Note swamp area formed between lateral moraine and hillside. Auger holes indicate 11 ft. of peat. Note also boulder in swamp with high water surface marks that show differential rotation. Slightly further down valley is end moraine. In addition to such features as previously reported, other end moraines have now been found at Noyes Pond, Pigeon Pond, Spring Lake, Lakota Lake, and Crook Brook indicating widespread Mountain glaciation in Vermont. Lunch, and then return to cars, proceed back down valley crossing Little River.
8.6	59.0	Join Route 100 north.
2.9	61.9	Stay on Route 100 through Stowe village.

Miles between Points	Cumulative Mileage	Description
3.2	65.1	Bear right leaving Route 100.
1.8	66.9	The first of a series of four deltas, some slightly pitted, that crest between 780 and 800 ft. These have been assigned to Lake Lamoille by Connally and to Lakes Gillett and Stowe by Wagner.
0.9	67.8	Road bends sharply left.
3.6	71.4	Sharp right turn ascending extensive 780 ft. delta deposited by upper Lamoille River.
1.3	72.7	Turn sharply back to left.
0.5	73.2	STOP 5 (Connally): From this vantage point the 780 ft. delta can be seen in the foreground and a partially collapsed or dissected 720 ft. delta can be seen in the distance at Hyde Park. In addition, small deltas are present from Morrisville to Johnson at 640 ft. The upper level is assigned to Lake Lamoille, the intermediate to Lake Mansfield, and the lowest to a Lake Coveville inlet. Wagner has assigned the upper level to Lake Gillett and the intermediate to Lake The Creek. We will discuss the relationship of the three levels to the Lake Gillett spillway. Continue toward Morrisville.
1.0	74.2	Morrisville, turn right on Route 100.
0.2	74.4	Cross Lamoille River.
0.9	75.3	Take Route 15 west.
4.0	79.3	A 740 ft. delta on the south edge of the village of Johnson.

Miles between Points	Cumulative Mileage	Description
0.7	80.0	Bear right on Route 100 in Johnson and continue north.
1.8	81.8	Another dissected 740 ft. delta just east of East Johnson.
1.2	83.0	An extensive delta that crests at 840 ft. was deposited here by the Gihon River.
2.0	85.0	Village of North Hyde Park.
2.7	87.7	Turn left on dirt road; note broad outwash surface.
1.5	89.2	STOP 6 (Wagner): Gravel pit in Phase I Mountain glaciation, Ritterbush Valley.
		Continue northward for 200 ft. and take dirt road to the left.
1.0	90.2	STOP 7 (Wagner): Ritterbush Pond; Phase II Mountain glaciation. Here we will examine the end moraines in Ritterbush Valley.
		Return to dirt road near Stop 6, turn left and continue northward.
1.0	91.2	View through trees to left of Ritterbush Pond cirque.
2.2	93.4	Enter Belvidere Pond cirque.
0.5	93.9	STOP 8 (Wagner): Scenic overlook and parking lot; Phase II Mountain glaciation. This is the Belvidere Pond cirque, "tarn", and end moraine.
		Continue west.
1.4	95.3	Junction Routes 109 and 118. Follow Route 109 south.
2.1	97.4	Gravel pit to left in Phase I Belvidere Valley Mountain glacier features.

Miles between Points	Cumulative Mileage	Description
1.0	98.4	Outwash plain(?).
0.5	98.9	Village of Belvidere Center.
2.1	101.0	STOP 9 (Connally): Pitted outwash is present almost certainly as a result of the Belvidere Pond glacier with possible additions from a local glacier immediately north of the stop. Although the surface elevation is only 800 ft. here it rises to 840 ft. to the north. Thus, Connally assigns this feature to Lake Lamoille, suggesting that local Mountain glaciation can be correlated with Glacial Lake Lamoille. Kettles are not present in Lake Mansfield deposits suggesting a very short-lived episode of local glaciation. Continue south.
3.8	104.8	Village of Waterville.
4.8	109.6	Junction with Route 108. Follow Route 108 south.
0.4	110.0	Junction with Route 15. Follow Route 15 west to Jeffersonville and from there to Burlington.
28.4	138.4	END OF TRIP.

Trip G-3

STRANDLINE FEATURES AND LATE PLEISTOCENE CHRONOLOGY OF NORTHWEST VERMONT

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Introduction

On this field trip we will examine early Holocene Champlain Sea strandlines along Lake Champlain; then we will see late Pleistocene glacial and glacial lake deposits that indicate both active and stagnant ice retreat in the northern Champlain Valley. Figure 1 is a location map indicating the area under consideration; Figure 2 is a map of the surficial geology of the Enosburg Falls quadrangle; Figure 3 is a north-south plot of features in the northeastern part of the Champlain Valley and adjacent Quebec including data from Wagner (this guidebook, Figure 3, p.322) and McDonald (1968).

Deglaciation of this region began with the retreat of the Laurentide ice sheet from the Green Mountains and Champlain Valley; in the latter there was apparently a lobe of ice which would persist in form as the ice retreated both northward and away from the Green Mountain front. Stewart and MacClintock (1969) discuss the first high-level proglacial lakes to form accompanying initial deglaciation. The presence of local mountain glaciation near Belvidere, Vermont (Wagner, 1970, 1971; Stewart, 1971; Connally, 1971) does not appear to influence deposits or events in the region under discussion here, other than being the source of outwash waters supplying sediment.

As deglaciation proceeded, large proglacial lakes gradually formed in the Champlain Valley at the ice margin, forming various stages of Glacial Lake Vermont, the two principal phases of which were the Coveville and Fort Ann phases, named for their presumed outlets in New York. Work done reported in this article confirms suggestions by McDonald (1968) and by Stewart and MacClintock (1969) that these water levels may have been confluent between the Champlain Valley and the area of southeastern Quebec studied by McDonald. It is proposed that the upper level, the "Sherbrooke phase" of Glacial Lake Memphremagog (McDonald, 1968), is at least in part correlative with a corresponding level in the Champlain Valley, probably the Coveville phase of Lake Vermont of Chapman (1937), and that the lower phase described by McDonald is correlative with the Fort Ann phase of Glacial Lake Vermont. Features to the south (see Wagner, this guidebook, Figure 3, p.322), corres-

ponding reasonably well to the Coveville level of Chapman, are traceable from the vicinity of Jeffersonville and Bakersfield, Vermont (at 720-740 ft.) north and northeastward up into the Missisquoi basin along the mountain front, and thence up the North Branch of the Missisquoi to the vicinity of Bolton Center, Quebec and the Lake Nick col (817 ft.) described by McDonald (1968, p. 668-669). Features belonging to a lower plane, approximately 120-140 feet below the first are likewise traceable from Jeffersonville and Bakersfield (600 ft.) up through the Missisquoi Basin and Sutton Valley to Lake Brome, Quebec. The data, when plotted on a north-south section, form two fairly well-defined curves (Figure 3). McDonald (1968) envisions the Cherry River moraine ice holding back the waters of the Sherbrooke phase of Glacial Lake Memphremagog, then the retreat of the ice beyond the Sutton Mountains, permitting the water level to drop to the lower level, possibly confluent with Fort Ann waters; he also notes (p. 692) that the lower lake system was already in existence at the ice front when the Highland Front moraine was developed. It should be noted that the northernmost point plotted in this article on the upper plane (Coveville-Sherbrooke), No. 20, is a well developed delta on the southeast flank of the Brome-Spruce-Pine mountain area, showing topset-foreset contact at 825 feet; this vicinity would have to be free of ice before Ft. Ann time. On the whole, the findings reported here agree with those of McDonald (1968).

Several things should be noted about Figure 3. First, all points are projected westward, and of necessity involve scatter due to the width of the area considered, and the fact that the isobases do not trend directly east-west. The curves appear to level off northward, as the locations gradually shift to the northeast, becoming more parallel to the isobars. Also, the elevations were determined using topographic maps and bench marks, and of necessity involve both variation and error.

Below the second curve there are a number of features which appear to represent levels intermediate to those of the Champlain Sea. These are best displayed in the Enosburg Falls quadrangle, between the towns of Enosburg and Bakersfield along Vt. Rt. 108, where a set of well defined multiple terraces can be seen (Points 4, 5, 24, 46-49 on Figure 3); these may correspond to intermediate phases between Glacial Lake Vermont and the Champlain Sea, "Lake New York" of Wagner (1969).

Marine waters entered the isostatically depressed Champlain Valley following retreat of the Laurentide ice mass from the St. Lawrence Valley. The oldest marine shell date in the Champlain Valley is from the marine shells at Stop 6, the gravel pit 2 miles south of Frelighsburg, Quebec, dated at $11,740 \pm 200$ years B.P. The highest marine strandline in the valley proper is straight and parallel to higher (older) proglacial lake water planes (Chapman, 1937; Wagner, 1972). Recent shell dates of lower (younger) marine shoreline deposits allow correlation of these features. Lower

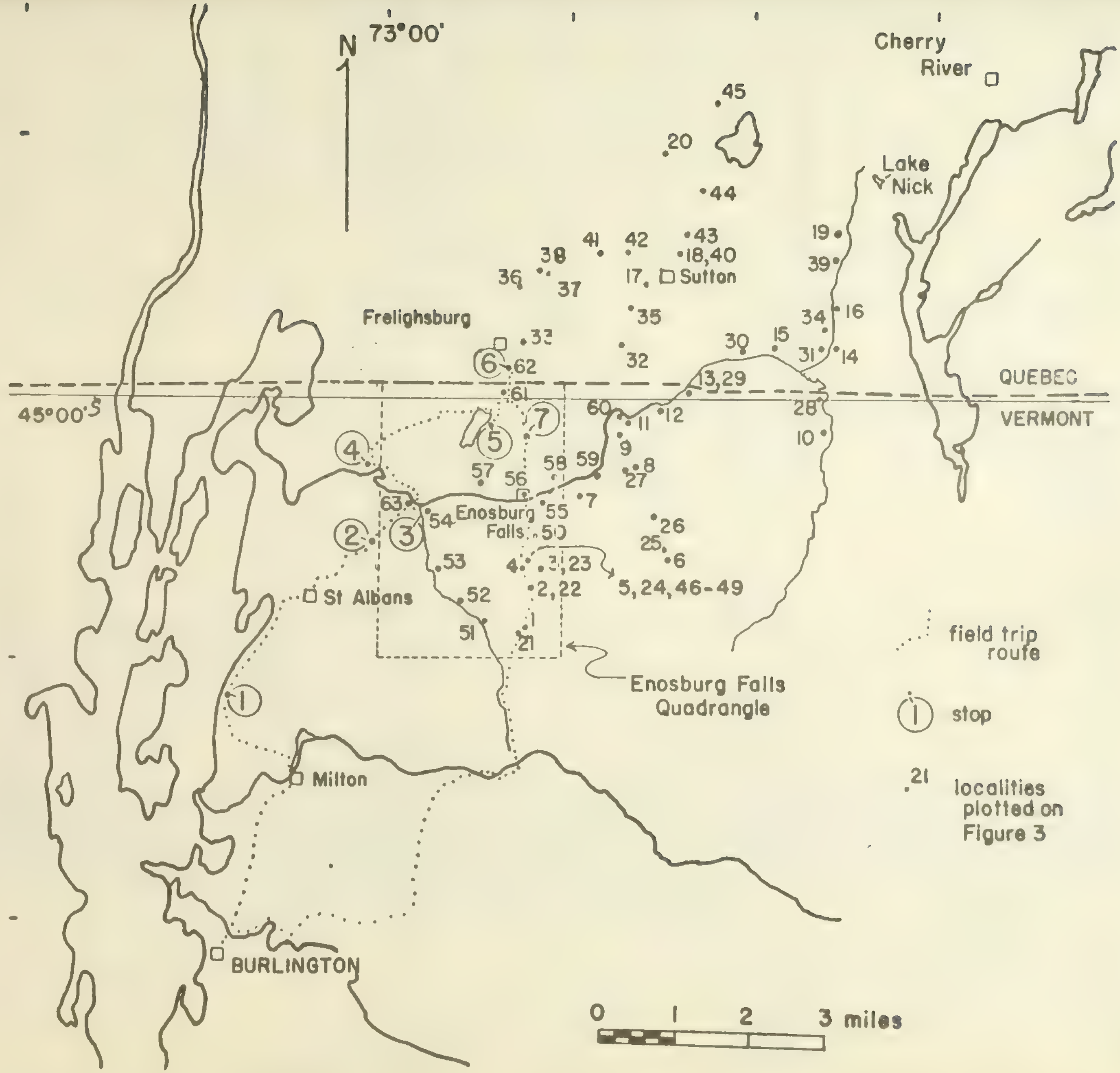
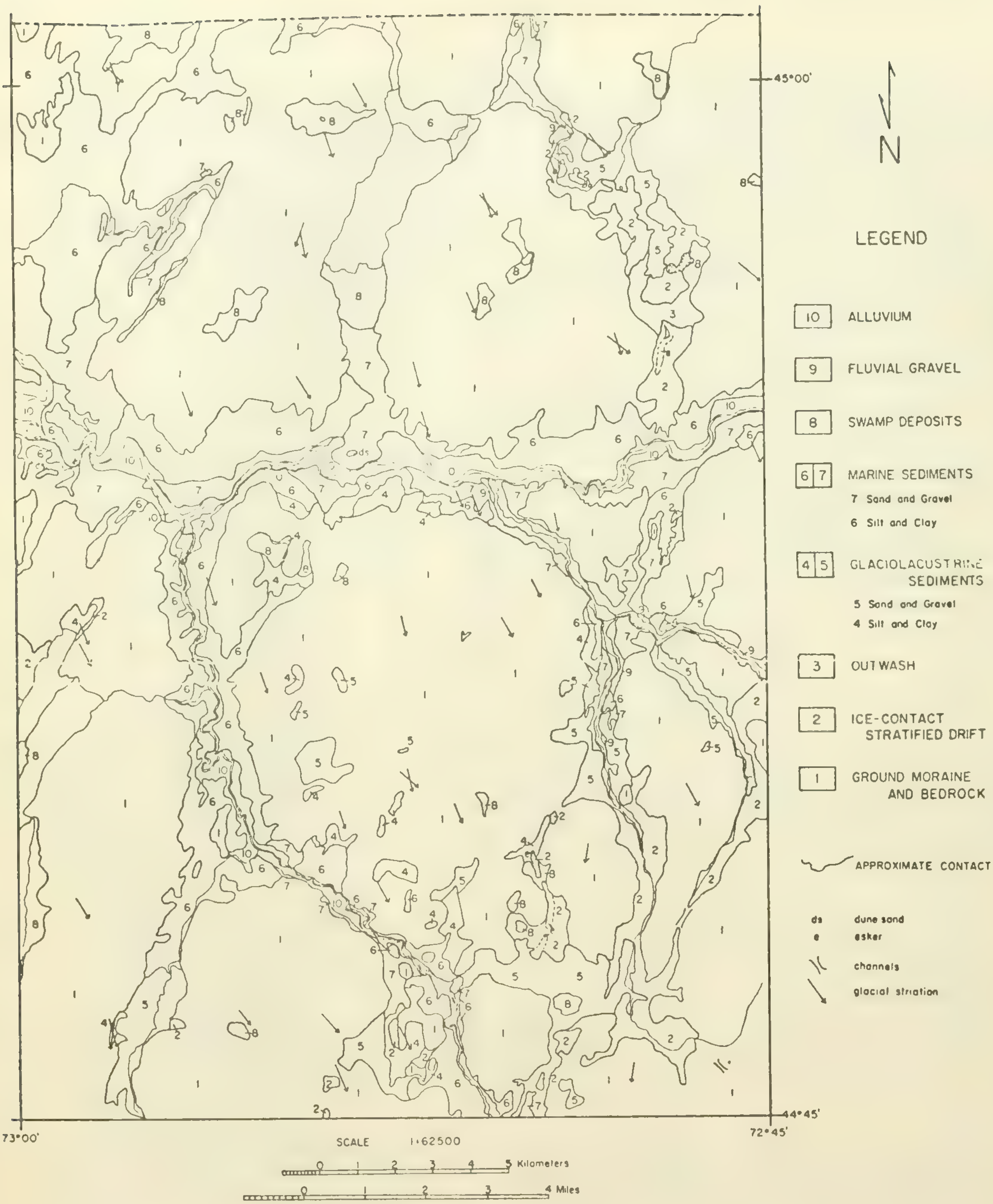


Figure 1. Location map of northwestern Vermont and adjacent southern Quebec.



SURFICIAL GEOLOGY OF THE ENOSBURG FALLS QUADRANGLE

Figure 2

W. R. PARROTT 1971

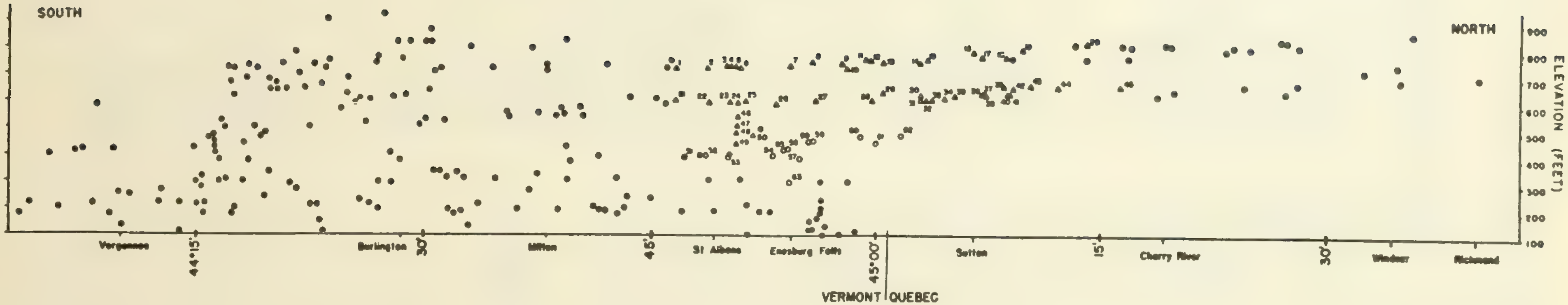


Figure 3. North-south section of shoreline features, northwestern Vermont and adjacent southern Quebec. Open figures pertain to this fieldtrip article (see Table 1 for locations); circles (closed or open) are from Wagner, (Fig. 3, p.), this guidebook; closed hexagons are from McDonald (1968). No data on lowest features in Quebec are shown. Locations are numbered sequentially from south to north.

TABLE 1

NUMBERED LOCALITIES PLOTTED ON FIGURE 3

Number	Feature	Town or Township	Elevation (feet)
1.	delta	Bakersfield, Vt.	720-740
2.	delta-terrace	Bakersfield, "	720-740
3.	beach deposits	Enosburg, "	740
4.	beach deposits	Enosburg, "	740
5.	beach deposits	Enosburg, "	740
6.	delta	Montgomery, "	720-740
7.	delta	Enosburg, "	740
8.	delta	Richford, "	760
9.	delta	Richford, "	740-760
10.	terrace	Troy, "	720-740
11.	delta	Richford, "	760
12.	delta	Richford, "	760
13.	delta	Richford, "	740-760
14.	delta	Potton, Que.	750
15.	delta	Potton, "	750-775
16.	delta	Potton, "	775-800
17.	delta	Sutton, "	750-775
18.	terrace	Sutton, "	750-775
19.	delta-terrace	Bolton, "	800
20.	delta	Brome, "	825
21.	delta-terrace	Bakersfield, Vt.	600
22.	terrace	Bakersfield, "	600
23.	beach deposits	Enosburg, "	600
24.	terrace	Enosburg, "	600
25.	terrace	Montgomery, "	600-620
26.	terrace	Montgomery, "	600
27.	terrace, delta	Richford, "	600-620
28.	beach deposits - terrace	Troy, "	600-620
29.	delta-terrace	Richford, "	640-660
30.	delta	Sutton, Que.	625
31.	terrace	Potton, "	600-625
32.	terrace	Potton, "	600-625
33.	terrace	St. Armand, Que.	600-625
34.	terrace	Potton, "	600-650
35.	terrace-(delta?)	Sutton, "	625
36.	terrace	Dunham, "	625-650
37.	terrace	Dunham, "	625-650
38.	terrace-(delta?)	Dunham, "	625
39.	terrace	Potton, "	650-675
40.	terrace	Sutton, "	625-650
41.	terrace (delta?)	Dunham, "	625-650
42.	delta?	Sutton, "	650

Number	Feature	Town or Township	Elevation
43.	terrace	Brome, Que.	650-675
44.	terrace	Brome, "	650
45.	terrace	Brome, "	650-675
46.			
46.	terrace	Enosburg, Vt.	550
47.	terrace	Enosburg, "	520
48.	terrace	Enosburg, "	490
49.	terrace	Enosburg, "	450
50.	terrace	Enosburg, "	480
51.	delta	Fairfield, "	380-400
52.	delta	Fairfield, "	400
53.	delta	Fairfield, "	380-400
54.	delta	Sheldon, "	400-420
55.	delta	Enosburg, "	420-440
56.	delta	Enosburg, "	440
57.	delta	Sheldon, "	380-400
58.	terrace	Enosburg, "	440
59.	delta	Berkshire, "	440-460
60.	delta	Richford, "	460-480
61.	delta	Berkshire, "	440-460
62.	delta	St. Armand, Que.	450-475
63.	delta	Sheldon, Vt.	300

strandlines show less tilt, indicating that crustal rebound began during the marine episode. Models and details of isostatic adjustment will be discussed at the first stop.

On Figure 3 a number of points related to the Champlain Sea have been plotted; however, data in the literature on Quebec are scanty and have not been included. It should be noted that the marine maximum of 540 feet noted by McDonald (1968, p. 673) fits the plot well and maintains parallelism with the upper two planes. In addition, the well developed features noted by McDonald at 400-420 feet appear to be characteristic of this area as well, although they vary up the Missisquoi basin from 380-400 feet in the west to 420-440 feet in the east.

The sequence of deglaciation affecting the influx and history of the Champlain Sea, however, now appears to be more complex than originally contemplated. Some workers (Cannon, 1964; Stewart and MacClintock, 1969) have proposed an intermediate period of subareal weathering between the Glacial Lake Vermont and the Champlain Sea; recent work by McDonald (1968), Johnson (1970), Wagner (personal communication) and the present study have not detected any weathering zone. However, mappings by Wagner in the St. Albans and Jay Peak Quadrangles, and by Parrott in the Enosburg Falls and Jay Peak Quadrangles have revealed the presence of a till within Champlain Sea sediments in the Missisquoi and Champlain Valleys. Shells found in the Frelighsburg, Quebec pit showing possible disturbance of the deltaic deposits there may record this readvance. The kame complex at Berkshire, Vermont is essentially surrounded by the till, but shows no evidence of disturbance itself, and appears to be related to the wasting away of the ice of this readvance. It is proposed that this readvance be tentatively named the "Missisquoi readvance," as the Missisquoi basin marks its apparent southern limit.

In all, deglaciation appears to have been at first characterized by active retreat, and ended in stagnation-zone retreat with the wasting of the Missisquoi ice.

Acknowledgements

We give special thanks to James Morse of the University of Vermont for assistance in the field and in the preparation of this article. We also gratefully acknowledge W. Philip Wagner for his advice and guidance in the course of the authors' research, as well as numerous other individuals who made the study possible. Gail Schwartz receives special thanks for typing the manuscript.

MAPS

Road Map

Vermont*

U.S.G.S.

<u>1:250, 000</u>

Lake Champlain NL 18-12*

<u>15 Minute Quadrangles</u>

Milton
St. Albans
Enosburg Falls*
Jay Peak
Irasburg
Mt. Mansfield

<u>7 1/2 Minute Quadrangles</u>

Milton
Georgia Plains*
St. Albans Bay*
St. Albans*
Highgate Center*

Canadian (obtainable from the Map Distribution Office
Department of Mines and Technical Surveys
Ottawa)

<u>1:500, 000</u>

Ottawa-Montreal NW 44/76

<u>1:250, 000</u>

Montreal 31-H

<u>1:50, 000</u>

Sutton 31 H/2 West
Sutton 31 H/2 East
Granby 31 H/7 East
Memphremagog 31 H/1 West

* <u>Suggested for field trip</u>

Road Log for Trip G-3

Trip will assemble in parking lot of University of Vermont College of Medicine, off of East Avenue. TRIP LEAVES AT 8:30 A.M. SHARP! AT STOPS, PARK CARS AS FAR OFF ROAD AS POSSIBLE.

Mileage

Cum.	S/S	
0	0	Road log begins. Leave College of Medicine parking lot and turn right onto East Avenue.
0.3	0.3	Intersection with U.S. Rt. 2; turn left onto Rt. 2.
0.8	0.5	Intersection with I-89 entrance; enter on right, after crossing over the interstate, heading north toward St. Albans.
1.9	1.1	Note marine delta sands in South Burlington landfill on right.
2.4	0.5	Note Winooski River gorge on right; 60 feet downcutting into dolostone bedrock since recession of Champlain Sea.
10.6	8.2	Milton Exit; turn left at end of exit ramp, proceed .1 mi. (E), turn left (N) on U.S. Rt. 7. ENTER MILTON 7 1/2' Quadrangle.
14.0	3.4	The southern end of the village of Milton sits on a maximum Champlain Sea delta (360' elevation). Note extensive delta flat north and northwest of the village.
16.7	2.7	Follow Rt. 7 through the village of Milton, noting unique pivot-gate dam at south end of Arrowhead Mountain Lake.
17.5	0.8	Lake Road. Turn left (W). Continue west and north. ENTER GEORGIA PLAINS 7 1/2' Quadrangle.
23.5	6.0	Bear left (W) on Lake Road.
24.8	1.3	You are crossing the Champlain Thrust contact: Dunham over Beldens(?); upper plate forms fault scarp along east shore of Lake Champlain.
25.0	0.2	Sharp right turn; proceed north to:
26.1	1.1	STOP 1. Champlain Sea beach deposit, 160' elevation, C-14 date 10,460 years B.P. on *Macoma balthica* shells. Note washed, imbricate structure of beach gravels.

Mileage

Cum.	S/S	
		What factors could influence the shell C-14 date here ? Discussion of models and details related to isostatic adjustment.
		Continue north along the lake. ENTER ST. ALBANS BAY 7 1/2' Quadrangle.
32.3	6.2	Melville Landing; turn right (SE).
32.8	0.5	Note delta (240' elev.) south of the road; this delta is above the projected 10,460 year-old strandline.
33.2	0.4	Turn left; Proceed north to Mill River.
35.2	2.0	Mill River; note downstream incision, well-developed floodplain and modern delta.
37.4	2.2	Proceed north to Vt. Rt. 36, St. Albans Bay; turn right.
38.2	0.8	Kellog Road; turn left (N). Is there beach topography along this road ? ENTER ST. ALBANS 7 1/2' Quadrangle.
41.9	3.7	Railroad crossing. Beach, wavecut topography on right.
42.1	0.2	Proceed 0.2 miles east to Route 7; turn right (S).
44.7	2.6	Intersection Vt. Rt. 105; turn left (E).
48.2	3.5	Intersection at Greens Corners; turn right (E).
48.6	0.4	Bear left at fork in road (NE).
49.8	1.2	STOP 2. Greens Corners delta and associated channel. Gravel is in kame delta or kame terrace formed when the Laurentide ice margin impinged against the upland. Till occurs at and near the top of the section in places. Similar till also is found at and near the land surface in the field to the west, and in many other localities north and northwest of this site. Northeast of this locality is a linear valley which now is occupied by a small stream. The gravel pit is at the divide, and is the northernmost location of Lake Greens Corners (Wagner, this guidebook). At least two other distinct channels, also approximately coincident with the water plane of

Mileage		
Cum.	S/S	
		Lake Greens Corners, occur in the upland area to the northwest. All of the above channels can be traced northeastward into the Missisquoi Basin where they extend to the level of deltas at the marine limit. Apparently, drainage from Lake Greens Corners was controlled by these channels in Champlain Sea time.
		Proceed northeast along channel and railroad tracks. ENTER ENOSBURG FALLS 15' Quadrangle.
53.2	3.4	Intersection with Vt. Rt. 105; turn right (E).
54.5	1.3	Intersection with road to Sheldon; turn right (S), onto it.
55.2	0.7	Sharp turn in road off to right (S); proceed straight ahead (E) onto dirt road, crossing: a. Black Creek bridge, and b. railroad.
55.5	0.3	STOP 3. Gravel pit in Champlain Sea deltaic material. Sand pit exposes deltaic sand of the marine limit Black Creek delta. In the eastern wall of the pit are exposed several feet of till overlying the deltaic materials. To the south the topset level of the delta is well-defined but the surface deposits are bottomset silt and clay. One well in the region penetrates through the silt-clay material which overlies sand. The delta is believed to be an early Champlain Sea feature with overlying lacustrine sediments from a temporary lake. The till is taken as evidence of glacial control for the lake.
		Return to Rt. 105 via route just taken.
56.5	1.0	Intersection with Rt. 105; turn right (E).
56.9	0.4	Cross Missisquoi River.
60.8	3.9	Intersection with Vermont Rt. 78; turn left (N). East Highgate; sharp turn to right (N). ENTER HIGHGATE CENTER 7 1/2' Quadrangle.
62.2	1.4	Intersection with small road off to left; turn left (SW).
63.1	0.9	STOP 4. Champlain Sea delta gravel pit; Sand and gravel pit on Champlain Sea sediments at the Port Kent level of Chapman (1937).

Mileage

Cum.	S/S	
		Return to intersection with Rt. 78.
64.0	0.9	Turn left (W) onto Rt. 78.
64.7	0.7	Beaulieus Corner; turn right (sharp), to northeast. RE-ENTER ENOSBURG FALLS QUADRANGLE.
66.8	2.1	You are now crossing a plain of Champlain Sea sediments, with several bedrock islands exposed.
67.7	0.9	Browns Corners. Proceed straight ahead (E). Stay on main road.
70.6	2.9	Franklin; intersection with Vt. Rt. 120. Turn left (N).
71.0	0.4	4 Corners; bear right, following Rt. 120 (E).
72.8	1.8	Lake Carmi. This lake rests in a valley containing only bedrock, till, and Champlain Sea sediments; now draining to the north, it originally drained southward following the Champlain Sea influx.
74.3	1.5	Intersection with road to right (on south); turn right.
77.4	3.1	STOP 5. LAKE CARMI STATE PARK. LUNCH. Time for discussion.
		Return to Rt. 120.
82.3	4.9	Intersection with Rt. 120; bear right (ahead) (N).
83.7	1.4	East Franklin; sharp turn to right (E).
84.2	0.5	Intersection with Vt. Rt. 108; turn left (N).
85.6	1.4	International Border; we will stop to be cleared as a group. Please refrain from having any items in your vehicle which might be a source of grief.
86.9	1.3	STOP 6. Gravel pit 2 miles south of Frelighsburg, Quebec. Refer to discussion above. Deltaic material here contained a lens of sand and clay containing disturbed *Macoma balthica* which dated at 11,740 ±200 years B.P.
		Return to Rt. 108; turn right (S).

Mileage

Cum.	S/S	
88.1	1.2	Re-cross International Boundary; we will stop to be let back into the United States.
89.5	1.4	Intersection of Rt. 120 with Rt. 108; bear left, toward West Berkshire (S).
90.1	0.6	Gravel pits in gorge to right contain fluvial gravels; This area drained an upland lake which we will see shortly.
90.3	0.2	West Berkshire.
90.5	0.2	Intersection with dirt road on right (S) side of Rt. 108. Turn right.
92.2	1.7	You are now driving through a kame field mantled by lacustrine sediments; drainage was down through the gorge we just came through.
92.8	0.6	Intersection with road between Berkshire and Enosburg Falls. Turn right (S).
92.9	0.1	The hills in front of you are part of the massive kame field in this area.
93.4	0.5	STOP 7. Gravel pit in Berkshire kame field. Kames in this area rise 200 feet above the surroundings in places and are quite extensive. This pit is the best exposure at the present time. Sediments show massive deposition of sands and gravels from stagnant melting ice, and are characterized by normal faulting. None, however, show evidence of thrusting or other disturbance as far as ice movement is concerned. To the southeast, south, and southwest, along the Missisquoi and Champlain Valleys, Champlain Sea sediments contain a till; this area shows no disturbance however: hence these deposits are interpreted as being post-Missisquoi readvance in age, probably related to stagnation of the ice of the readvance.

- -

This stop officially concludes the field trip. The route southward suggested below, via Vt. Rts. 108 and 15, passes through Enosburg Falls, Jeffersonville, and Cambridge.

The following log incorporates several features of significance along the return route:

Mileage

Cum.	S/S	
93.5	0.1	Turn right as you leave the gravel pit, and proceed south.
96.6	3.1	Intersection with Vt. Rt. 105. Turn right onto Rt. 105 (W).
97.1	0.5	Enosburg Falls, elevation 422 ft. resting on Champlain Sea deltaic sediments.
97.4	0.3	Intersection where Rt. 105 bears right (W), and Rt. 108 continues straight ahead (S). Proceed straight ahead.
100.0	2.6	West Enosburg; From here on, for the next 4 miles, the valley of Tyler Branch and The Branch contains at least 6 levels (see Figure 3, points 4, 5, 24, 46-49), from the upper lacustrine ("Coveville") to the Champlain Sea. These are clearly visible as you drive along the length of the valley ahead.
104.7	4.7	Browns Pond. Above you to the left is a kame terrace which extends southward, merging into the Bakersfield delta at 740', later reworked to 600' during "Ft. Ann."
106.7	2.0	Bakersfield. The town is built upon the 740' surface.
107.5	0.8	On southern side of valley, if you turn and look back to the right (NW) you can see the 600 foot surface in deltaic deposits south of Bakersfield. ENTER MT. MANSFIELD 15' Quadrangle.
110.7	3.2	Off to right, in the floor of the valley with a small farm resting on it, is a small moraine containing cobbles and Champlain Sea sediment, and which may mark the southern limit of the readvance; here a minor lobe extended down the Missisquoi Black Creek Valley from the northwest.
112.7	2.0	You are now in the Black Creek Valley, which was a channel southward at one time for glaciofluvial waters; the glaciofluvial sediments are mantled first, by lacustrine sediments, and then by Champlain Sea sediments.
118.0	5.3	Intersection with Vt. Rt. 109; bear right, staying on Rt. 108.
119.5	1.5	Intersection with Vt. Rt. 15. Turn right (W),

Mileage		
Cum.	S/S	
		toward Cambridge.
121.9	2.4	Cambridge bridge; turn left (S), across bridge, following Rt. 15 into the town of Cambridge.
122.9	1.0	Intersection with Vt. Rt. 104.

2 choices:

1. Bear left, and follow Rt. 15 through Underhill, Essex Center, and Essex Junction to Interstate 89 and Burlington.

2. Proceed straight ahead along Rt. 104 to Fairfax, and turn left (W) onto 104A, 2 miles beyond toward Milton, along the Lamoille River and Arrowhead Mountain Lake; turn right onto U.S. Rt. 7 at intersection with Rt. 9, then onto I-89, and south toward Burlington and points beyond.

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Trip G-5

TILL STUDIES, SHELBURNE VERMONT

by

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Department of Geology, University of Vermont

INTRODUCTION

In the summary report of the Vermont Geological Survey-sponsored surficial geology mapping program, Stewart and MacClintock (1969) presented the first comprehensive Laurentide stratigraphy for the entire state. Surface tills in three regions are differentiated primarily on the basis of till fabric. In a streambank exposure near Shelburne village in northwestern Vermont their "Burlington till"(northwest fabric) is reported overlying "Shelburne till"(northeast fabric)(Figure 1). This locality is the subject of this report. The report emphasizes till fabric measurements but other parameters are included: color, texture, lithology, particle shape, heavy minerals, and striae. The bulk of the data is from the exposure previously studied by Stewart and MacClintock but nearby exposures were also sampled.

Acknowledgements

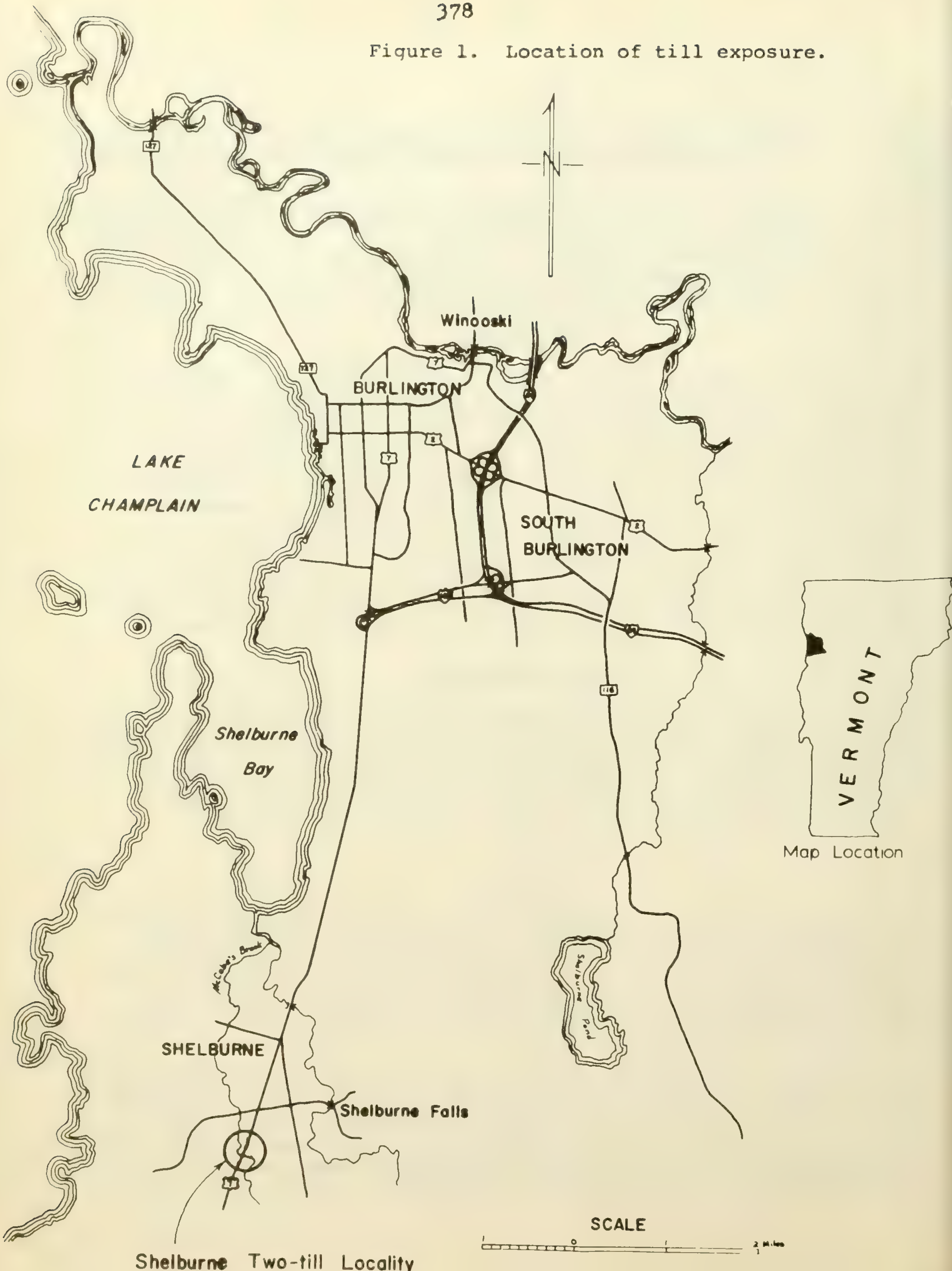
Preliminary till fabric measurements at the Shelburne locality were made in 1966 by M. W. Hebb and S. J. Minor, and in 1969 by C. A. Howard, Jr. and W. R. Parrott, all students at the University of Vermont. The bulk of the data presented here was collected in 1969 by the senior author with the assistance of B. P. Sargent and R. Switzer, also students at the University.

TILL COLOR

The first study of the Shelburne locality was made by Stewart (1961, p. 102) who reported northeast fabric maxima in the lower, gray-colored part of the till and northwest maxima in the upper, brown-colored part. Although he indicated that till color differences do not necessarily have stratigraphic significance, color is clearly used as a basis for till differentiation (Stewart, 1961, Figure 2 and p. 102).

A different interpretation of the till colors at Shelburne was presented by Thomas (1964) who believed that the brown coloration is due to oxidation of gray colored till. As evidence, he cited the pronounced weathered character of particles in the brown unit compared to the gray unit, and similar color variations in ponded silt and clay deposits in the area.

Figure 1. Location of till exposure.



From our study of the Shelburne exposure the following observations can be made about the color difference. Munsell color codings of wet samples are 10 YR 2/1 and 10 YR 3/4 for the gray- and brown-colored till units respectively, with relatively slight intra-unit variations. The contact between the brown and gray colors is sharp. Lenses of gray-colored till are surrounded by brown-colored till, and brown coloration extends downward along joints for several feet into the gray-colored unit. In this and numerous other exposures in the region showing similar brown-and gray-colored till, the contact between the two units generally appears to reflect the slope of the overlying ground surface. These aspects indicate to us that the color difference can be better explained by weathering, as Thomas suggested, than by multiple glaciation. Furthermore, our attempts to differentiate the brown- and gray-colored tills with a variety of parameters, including fabric, have been unsuccessful, thereby lending support to the view that multiple glaciation is not the cause of the color difference.

FABRIC

Stewart (1961, p. 102) reported fabric maxima (based on a 180 degree, two-dimensional reference system) of N30E and N15W for the gray- and brown-colored tills, respectively. Thomas' (1964, p. 68-72) fabric study of the same exposure showed N25W and N45W maxima for gray and brown units, respectively. He also measured fabric at a nearby exposure of gray-and brown-colored till, both of which had preferred north-south orientations. In unpublished fabric studies at the same locality, students from the University of Vermont consistently have found northeast maxima in the gray-colored till; in the brown-colored till, on the other hand, most fabrics showed bimodal distributions with northeast and northwest concentrations of varying relative strengths.

In the work reported here ten fabric analyses were made at the Shelburne locality (Figure 2). Eight of the sites were from two vertical trenches excavated to assure undisturbed samples, and two sites were at the middle and upper central parts of the exposure. A hand-held Brunton compass was aligned parallel to long axes of elongate particles to measure azimuth and inclination. In addition, the orientation of blade-and disk-shaped particles was determined by measuring the strike and dip of a plexiglass plate oriented parallel to flat particle sides. Thus, only azimuth and inclination were measured for rod -shaped particles, only strike and dip for disk-shaped particles, but both spatial factors were measured for blade-shaped particles. Long axis measurements are probably accurate to $\pm$ 5 degrees, whereas strike-dip data are somewhat less accurate.

The data were originally plotted in the field on Schmidt equal-area stereo nets. This showed that most fabric patterns are polymodal, thus making statistical reduction difficult. To facil-

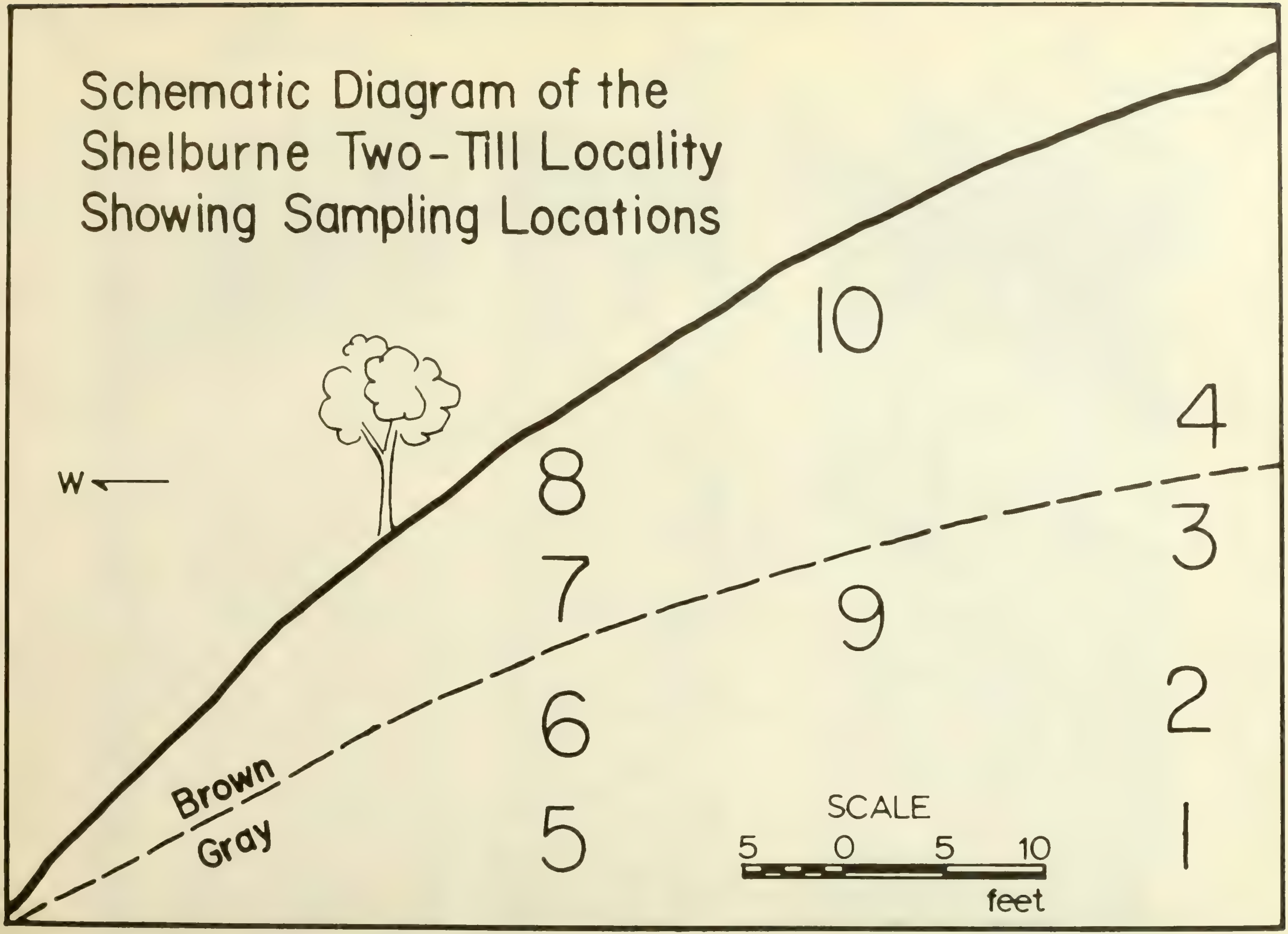


Figure 2. Schematic view of till exposure.

itate reduction of data, the computer program by Spencer and Clabaugh (1967) was used. Computer printout for long axis data is reproduced in Figure 3, and for pole plots of flat particle data in Figure 4. Long axis fabrics vary from sample to sample, but most tend to be characterized by azimuth maxima predominantly in the northeast and southwest quadrants. Thus, no correlation between fabric and color difference is apparent. Figure 3 also suggests that the inclinations of long axes are not significantly different from horizontal. This has been further established by simple statistical analyses. Sample 8 is unique for its predominant northwest - southeast concentration. Due to its proximity to the land surface, it is thought that the till in the vicinity of sample 8 might be disturbed by mass movement.

In Illinois Harrison (1957a) found that flat particles tended to dip in the upglacier direction. Krumbein (1939, Figure 3) depicted short axis plots (comparable to the flat particle fabrics reported here) arranged in a girdle oriented perpendicular to the flow direction. Flat particle fabrics from the Shelburne locality vary considerably but there is a suggestion that the strike of flat particles tends to parallel long axis trend of elongate particles, similar to Krumbein's findings. Note the similarities between strike of flat particles (Figure 4) and trend of elongate particles (Figure 3) for samples 2, 4, 7, and 8. Although not enough is known about flat particle fabrics, it appears that such measurements may lend support to long axis data.

About 200 feet south of the major exposure is a streambank showing gray-and brown-colored till. Three till fabric samples from this exposure all have strong northeast-southwest maxima.

Directly across the stream from the major till bank is a small exposure of gray till directly overlying bedrock. Additional fabric measurements were made by sampling from two vertical working faces oriented perpendicularly to each other and from a third, horizontal face perpendicular to the other two. Thus, fabrics taken from the same till, but from working faces of different orientations, can be compared. The apparent influence of working-face orientation on long axis fabrics is striking (Figure 5). Poles to working faces are represented by circles on the fabric diagrams in Figures 5A and 5B. It is believed that working-face orientation can introduce a significant bias in some cases due to a tendency to oversample particles projecting at high angles to the working face. Although conscious efforts were made to avoid such a bias, the relative difficulty experienced in extracting particles oriented nearly parallel with any working face made this impossible. Because the majority of till stones plunge at low angles, a horizontal working face might introduce less bias than other working faces. If this is the case, then the fabric of this cube of till is most likely northwest, as the diagram from the horizontal working face indicates. Such a trend is exaggerated by the working face oriented N70E. For the N20W working face,

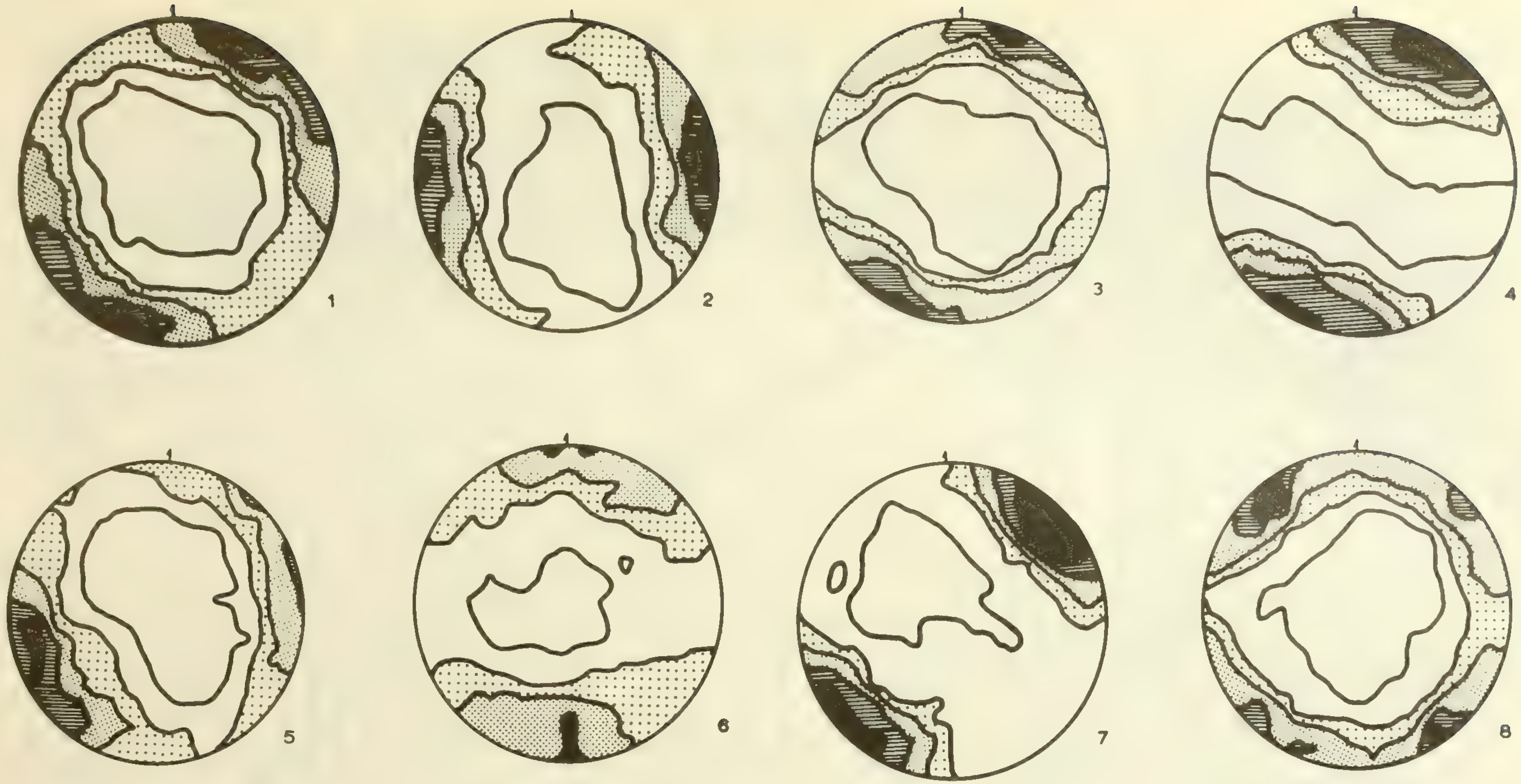
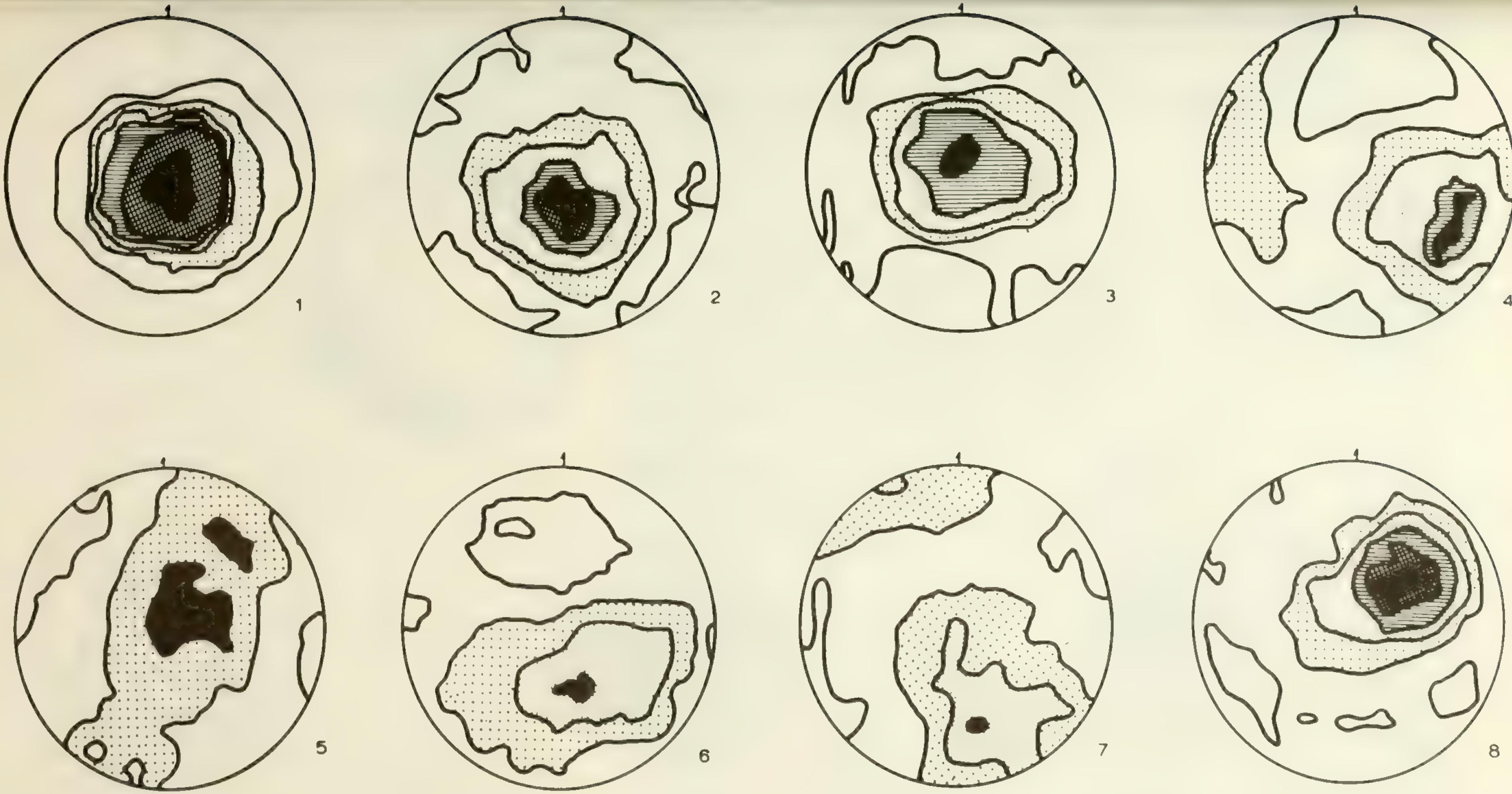


Figure 3. Lower hemisphere equal area diagrams for trend-plunge of long axes. Numbers refer to the sampling locations shown in Figure 2. Contour intervals and sample size for each diagram are: 1) 1,3,5,7,8 standard deviations (sigma), n=55; 2) 1,3,5,7,9 sigma, n=40; 3) 1,3,5,7,8 sigma, n=49; 4) 1,3,5,7,9 sigma, n=31; 5) 1,3,5,7,8 sigma, n=38; 6) 1,3,5,7 sigma, n=32; 7) 1,3,5,7,9,10 sigma, n=43; 8) 1,3,5,7,8 sigma, n=74; 9) 1,3,5,7,9 sigma, n=38; 10) 1,3,5,7,9,11 sigma, n=63.



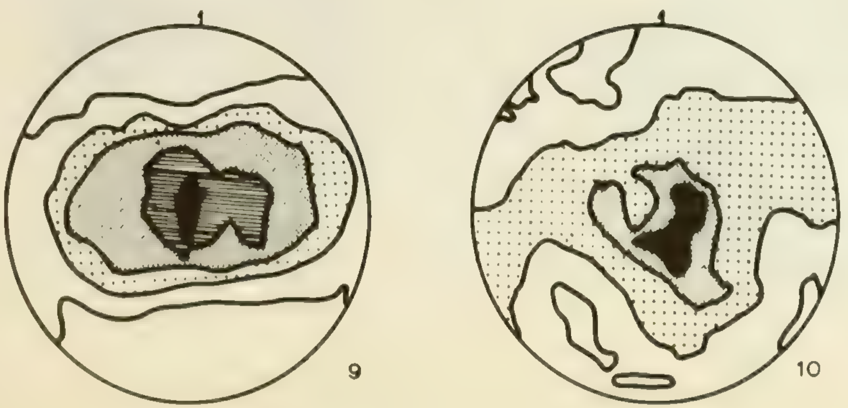


Figure 4. Lower hemisphere equal area diagrams for strike-dip of flat particles. Numbers refer to the sampling locations shown in Figure 2. Contour intervals and sample size for each diagram are: 1) 1,3,5,7,9,11 standard deviations (sigma), n=29; 2) 1,3,5,7,9,10 sigma, n=47; 3) 1,3,5,7,9 sigma, n=34; 4) 1,3,5,7,8 sigma, n=30; 5) 1,3,5 sigma, n=23; 6) 1,3,5,7 sigma, n=44; 7) 1,3,5,7, sigma, n=29; 8) 1,3,5,7,9,10 sigma, n=57; 9) 1,3,5,7,9 sigma, n=34; 10) 1,3,5,6 sigma, n=69.

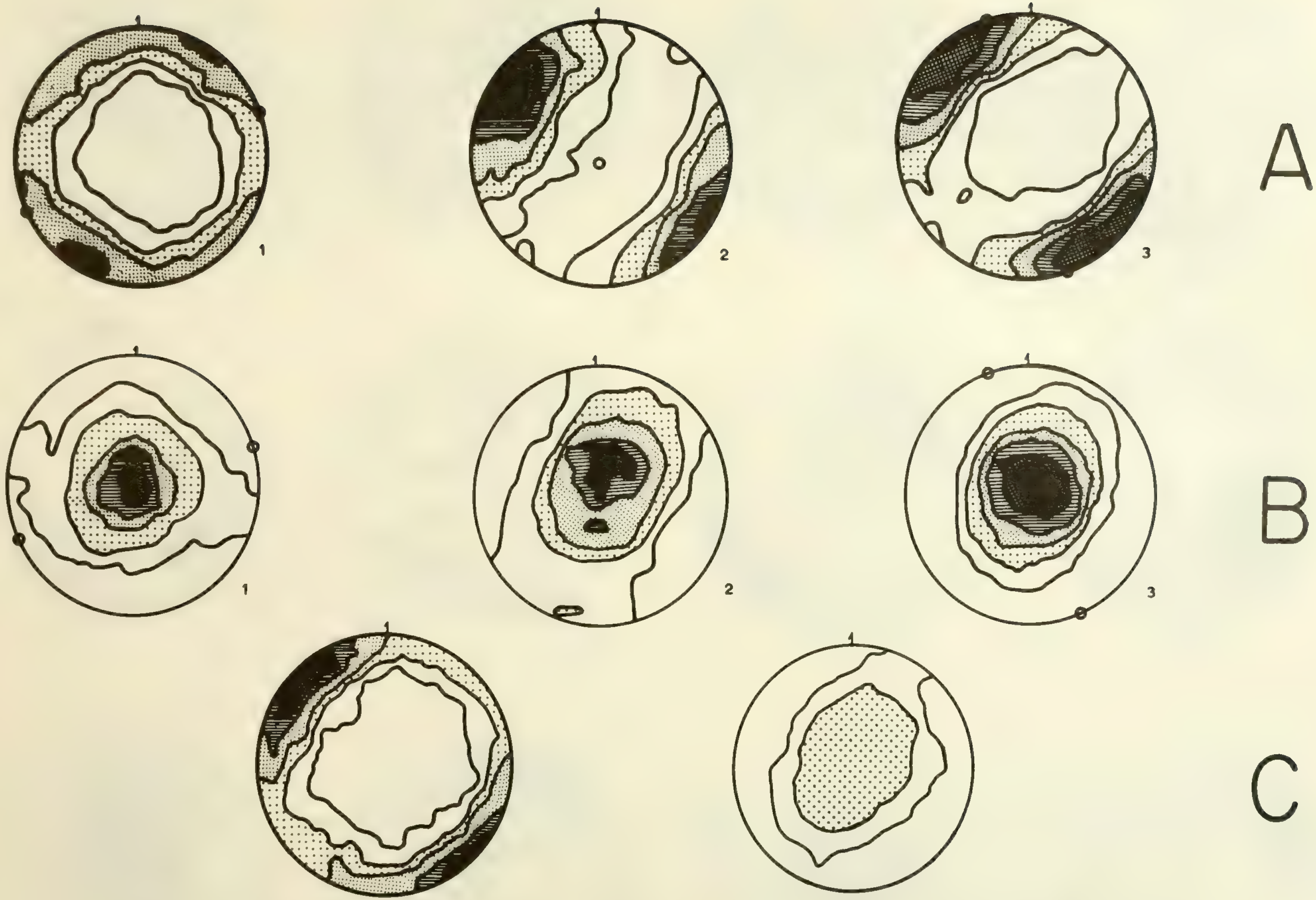


Figure 5. Lower hemisphere equal area diagrams for samples taken from the small exposure across the stream from the major till bank. The circles on the diagrams in 5(A) and 5(B) represent poles to the working face. (A) Trend-plunge of long axes. Contour intervals and sample size for each diagram are: 1) 1,3,5,7 standard deviations (sigma), n=42; 2) 1,3,5,7,9 sigma, n=25; 3) 1,3,5,7,9,11 sigma, n=39. (B) Strike-dip of flat particles. Contour intervals and sample size for each diagram are: 1) 1,3,5, 7,9,10 sigma, n=67; 2) 1,3,5,7,8 sigma, n=26; 3) 1,3,5,7,9,10 sigma, n=28. (C) Left: trend-plunge composite diagram combining all data from (A). Contour intervals: 1,3,5,7,8 sigma, n=106. Right: Strike-dip composite diagram [illegible]

on the other hand, it appears that a dominant but artificial northeast-southwest maximum is created with a lesser concentration in the northwest-southeast quadrants. Such a pattern is extremely misleading. Similar findings have been reported by Johansson (1968, p. 206), Dreimanis (1959), and Thomas (1964), all of whom recognized sampling bias as a significant problem. Although some of our fabrics were taken on non-horizontal working faces, most were taken on horizontal working faces, which are probably less subject to this type of bias.

To what extent till fabric measurements are reliable is problematical, but the dominantly northeast-southwest mode is at least internally consistent. It is our view that stratigraphic subdivision of the Shelburne exposures is not supported by reproducible fabric data.

TEXTURE

No thorough investigation of till texture was made in this study. Thomas (1964, p. 135) reported similar grain size distributions in both gray-and brown-colored till except that the brown till contained 10 percent clay versus 25 percent in the gray till. Kodl (1967) studied the -2Ø to greater than 4Ø size range and found no significant differences between a total of six samples of gray-and brown-colored till. In our work, caliper measurements of long, intermediate, and short axes were made of all particles used in the fabric analysis. Visual inspection of graphs of frequency of various particle sizes showed that all samples appeared similar except that samples 7 and 8 contain a higher proportion of particles with large long axes than other samples.

LITHOLOGY

Thomas (1964, p. 165) measurements of lithology of the granule fraction showed no significant differences between the gray- and brown-colored tills. Our work included identification of the lithology of particles selected for fabric analyses. This data shows considerable variations in particle lithology percentages between samples. However, no trends or patterns are apparent.

Coarse Fraction Particle Shape: Thomas (1964, p. 76-89) compared the shapes of carbonate and shale granules in the two tills, concluding that there was no significant difference. The coarse fraction ($\gtrsim$ 3.3Ø) axial measurements from this study were plotted on shape triangles as described by Folk (1964). It should be pointed out that inasmuch as only those particles suitable for fabric measurements were considered, equant grains were excluded. Nevertheless, the procedure used was similar in all cases so that the approach is at least internally consistent. Over 90% of the particles can be circumscribed by 0.2 - 0.5 short:

long axial ratio values and by 0.1 - 1.0 long-minus intermediate: long-minus short axial ratio values. A notable exception to this is sample number 8 which includes a larger proportion of compact particles than any other sample. It appears that this anomaly cannot be explained lithologically and is believed to be a significant, difference between samples.

Heavy Mineral Analysis: Heavy mineral separations of the medium sand-sized fraction were conducted by Caldwell (1969) on single samples from each of the brown and grey tills. Using bromoform (sp.gr.$\geq$ 2.85), amounts of 81.4% and 77.2% dolomite were counted in the heavy fraction of brown and gray till samples, respectively. Although the dolomite contents do not appear significantly different, Caldwell noted that dolomite in the brown till had a "pinkish-orange" color thought to be due to oxidation of the brown till. Removal of the dolomite by iodide-acetone separation (sp.gr.$\gtrsim$ 3.0), facilitated identification of the non-dolomite fraction, but no pronounced differences between the tills were found. Also, no significant differences of magnetic fractions were detected.

Clay Fraction Mineralogy: A preliminary X-ray diffraction analysis of the clay fraction was made by Parrott (1968). Sixteen samples spanning the contact between the gray- and brown-colored tills were taken to determine if any differences between the tills could be detected. Parrott found a deficiency of calcite near the top of the exposure, and a calcite concentration at slightly greater depth, both of which he attributed to leaching. Chlorite generally decreases upward, and muscovite, montmorillonite, and an unidentified mixed-layer mineral increase upward. Kaolinite content is lowest just above the color contact. Plagioclase is highest at the middle of the exposure, near the contact. Parrott (1968) concluded that no systematic differences between gray- and brown-colored tills could be found on the basis of clay mineralogy.

STRIAE

Glacial striae are well developed on a bedrock surface immediately south of the main till exposure and adjacent to the site where the data for the fabric diagrams shown in Figure 5 were collected. Two directions of striae are discernable. Over most of the outcrop surface only one set of striae are found with a N30W-S30E trend. In the most recently exposed part of the bedrock another set of striae oriented north-south appear. It is difficult to determine the relative age of the striae. The significance of the striae in relation to the main till exposure is unknown.

SUMMARY AND CONCLUSIONS

Our work on till fabric at the Shelburne - Burlington till locality does not support the two till view. Three independent

investigations of till fabric at the site have produced different results. Moreover, we have failed to find evidence of stratigraphic difference on the basis of a variety of parameters. Drawing an analogy from the null hypothesis of statistics, no significant stratigraphic differences at Shelburne can be inferred until conclusively proven. We prefer to avoid the usage of the terms "brown till" and "gray till" in deference to adjectival expressions without stratigraphic overtones, i.e. gray- or brown-colored till.

Perhaps the most worthwhile information from this study relates to the subject of till fabrics in general. The suggested bias due to working face orientation indicates that the concepts of transverse and longitudinal fabric maxima may not be straight forward. For example, a vertical working face oriented nearly parallel to the direction of former ice movement might result in a tendency to undersample the longitudinal population while over-sampling the transverse (refer again to Figure 5). Whole-till sampling techniques, as for example the method of Harrison (1957b), may be less misleading than our method of fabric measurement.

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Trip G-6

WOODFORDIAN GLACIAL HISTORY OF THE CHAMPLAIN LOWLAND, BURLINGTON TO BRANDON, VERMONT

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INTRODUCTION

The surficial geology of the Champlain lowland and bordering Green Mountains of west-central Vermont has been generally known for many years. However, the results of a recent, comprehensive, state-wide study in Vermont and studies in the upper Hudson Valley of New York have led to the definition of one major problem and the reinterpretation of two significant aspects of deglaciation; the physical characteristics of the waning glacier and the extent of proglacial lakes impounded by the retreating ice margin.

At least two tills are present in the Champlain Valley in western Vermont and one in the Connecticut Valley in eastern Vermont. The problem is whether the eastern till correlates with the upper or the lower Champlain Valley till.

The many kame terraces which flank the Green Mountain front throughout the Champlain Valley were incorrectly correlated by early workers. This led to the conclusion that the last glacier had stagnated and downwasted in place. However, it has now been shown that each kame terrace belongs to a discrete south-sloping sequence of ice-contact and outwash deposits. The sequences formed successively during recession of the margin of a still-active glacier.

Early workers correctly concluded that the clays, sands, beaches, and deltas flanking the Green Mountains were results of proglacial lakes. They inferred that these lakes were confined to the Champlain Valley. It has now been shown that the highest levels in the Champlain Valley were coextensive with similar lakes in the Hudson Valley which has led to an updating of terminology.

ACKNOWLEDGEMENTS

Field study by Calkin (Middlebury Quad.) and by Connally (Brandon and Ticonderoga Quads.) was supported by the Vermont Geological Survey. We are indebted to Dr. Charles G. Doll, State Geologist for his help, and to Dr. David P. Stewart of Miami University for many stimulating discussions and introduction to the field areas.

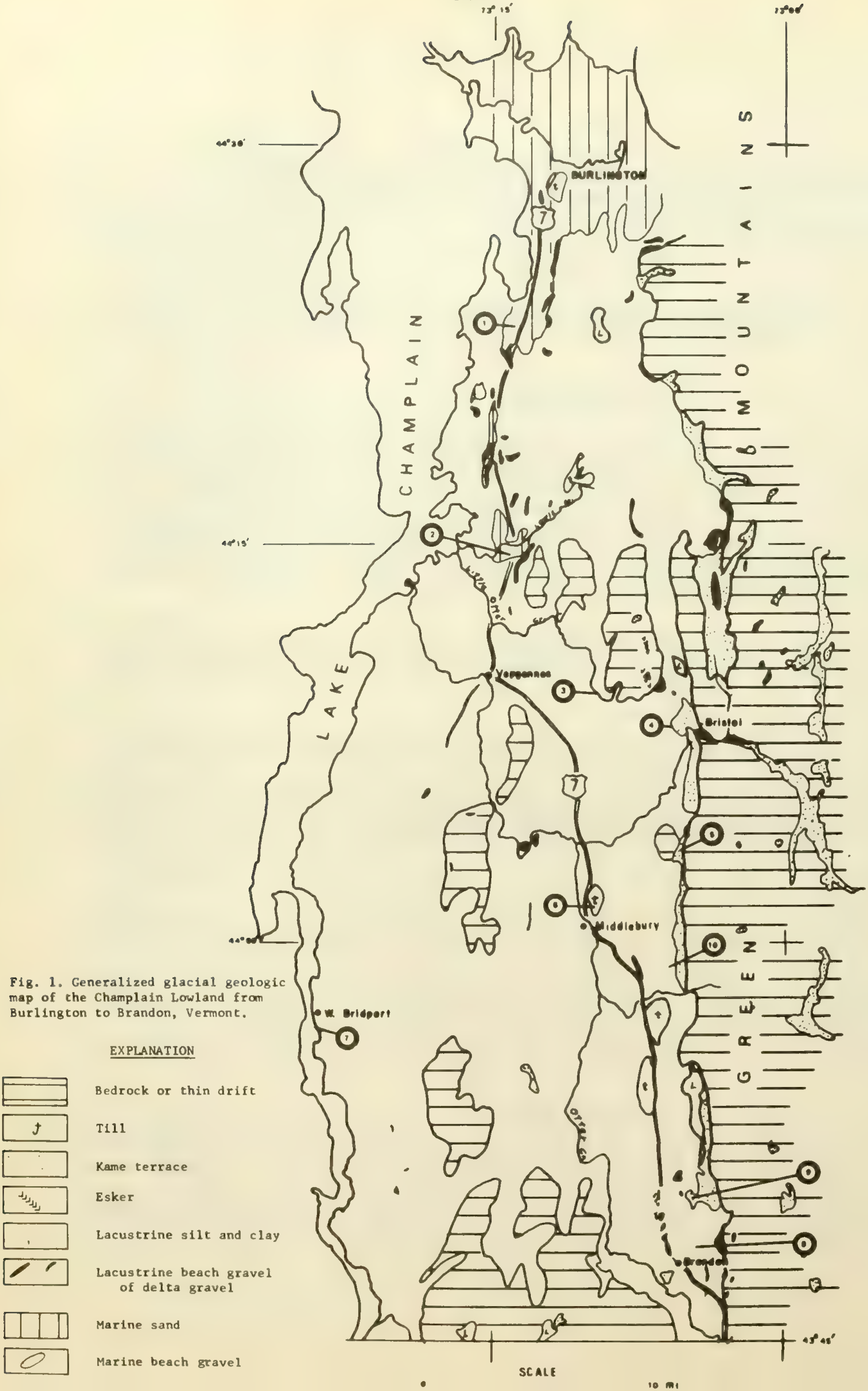

Fig. 1. Generalized glacial geologic map of the Champlain Lowland from Burlington to Brandon, Vermont.

EXPLANATION

Bedrock or thin drift

Till

Kame terrace

Esker

Lacustrine silt and clay

Lacustrine beach gravel of delta gravel

Marine sand

Marine beach gravel

THE BURLINGTON-SHELBURNE PROBLEM

The Champlain lowland and bordering Green Mountains have been overridden at least twice and probably three or more times by continental ice sheets during the Pleistocene. McDonald and Shilts (1971) record at least three distinct glaciations in Quebec, to the north, while Borns and Calkin (1970) distinguish at least two in northwestern Maine, to the east. Local evidence includes multiple-till sections, differences in till fabric orientations, and striations on scoured bedrock surfaces, all reported by Stewart and MacClintock (1969).

In four well exposed, multiple-till sections (Shelburne, Lewis Creek, Little Creek and West Bridport sites (see stops 1, 3 and 7, Figure 1) between Burlington and Brandon, lodgement tills with northwest fabrics are underlain by similarly compact tills with slightly different lithologies and northeast fabrics. At the Lewis Creek and Little Otter Creek sites varved clay records an ice recession between deposition of the contrasting tills. The observations of a northwest-derived surface till over a northeast-derived till is supported at several places in the Middlebury, Brandon, and Ticonderoga quadrangles where weak but definite northeast striae are cut by northwest striae. Supporting evidence from fabrics and striae is reported by Stewart and MacClintock (1969) for the bordering mountainous areas of northwestern Vermont . Although the division unfortunately has been based almost entirely on till fabrics, and there are occurrences of apparently contradictory till fabrics; the evidence for two till sheets in west-central Vermont is convincing.

Stewart and MacClintock (1969) defined the lower, northeast-derived till as the Shelburne till and the upper, northwest-derived till as the Burlington till. Some workers have questioned the interpretation of two tills at the type section of the Shelburne till, but a more important Burlington-Shelburne problem, discussed by Stewart and MacClintock (1969, p. 190), arises relative to the definition of the boundary between these tills and the correlation of the lower, northeast-derived lodgement till of the Champlain Valley with an ablation till with northeast fabric in eastern Vermont. The sandy ablation deposits of southeastern Vermont may well be the result of normal reorganization and lobation of a thinning ice mass in the north-northeast-trending Connecticut Valley; therefore, these could have been laid down by the same continental glacier that deposited the Burlington till in northwestern Vermont as suggested by Shilts and Behling (1967) and postulated by Stewart and MacClintock (1969, p. 80) as their Alternate Hypothesis III.

However, the Burlington-Shelburne problem is resolved, the Burlington appears to be a lithologically correlatable till sheet in northwestern Vermont. The Burlington till may represent the late

Woodfordian Luzerne readvance; the underlying Shelburne till deposited by the ice sheet that receded from the main Woodfordian Ronkonkoma Moraine on Long Island about 18,000 yrs. B. P. (Connally and Sirkin, 1972). Alternately, the Burlington drift may be a western facies of a much more extensive drift sheet that represents the entire Woodfordian. In either case it is possible that at least some of the lower tills of multiple till sites in the Champlain Valley record pre-Woodfordian glaciation. However, Connally (1970) postulated that both lodgement tills at the West Bridport section are from the last Woodfordian glaciation because of the orientation of striae on the smoothly polished bedrock surface beneath the till.

ACTIVE ICE RETREAT

Recession from the Burlington glaciation involved stagnation and downwasting in the Green Mountains while backwasting of an active, calving, ice margin occurred in the Champlain lowland where the terminus fronted a series of expanding glacial lakes. The wide, and apparently continuous series of kame terraces depicted in Figure 1 can be separated into discrete sequences. Each sequence grades southward from ice-contact deposits, through kame moraines, and onto outwash aprons. Connally (1970) describes five separate sequences in the Brandon quadrangle, each of which includes one or more kame terraces. Elsewhere, the presence of interbedded tills and lacustrine deposits in numerous subsurface exposures indicates that the recession of the Burlington glacier involved frequent frontal oscillations (Calkin, 1965).

Calkin (1965) demonstrated that large remnants of stagnant ice downwasted in depressions in the Green Mountains producing an abundance and variety of dead-ice deposits while continental ice was still actively receding in the Champlain Valley. These upland remnants shed outwash down the major valleys from high mountain divides and onto the retreating ice sheet. The outwash forms the bulk of the ice-contact drift in many of the kame terraces adjacent to the Green Mountains.

Connally and Sirkin (1970) suggest that the Burlington drift of Vermont is equivalent to the till of the Luzerne readvance near Glens Falls, New York and is therefore about 13,200 years old. Recession from the Luzerne readvance was underway by 13,150 yrs. B.P. The ice sheet retreated steadily northward through the Champlain Valley interrupted only by the Bridport readvance (Connally, 1970). This readvance extended from the vicinity of Burlington to near Bridport about 12,800 yrs. B.P. (Connally and Sirkin, 1972). No moraine marks the terminus of this readvance; glacial lake waters apparently prevented formation of any distinct recessional moraines in the lowland. Overriding of lacustrine deposits and calving of the active ice margin of the Bridport readvance probably caused the ubiquitous bouldery clay shown by Stewart and MacClintock (1970) between Burlington and Bridport.

GLACIAL LAKE HISTORY

Chapman (1937, 1942) made an exhaustive study of lacustrine and marine strandlines in the Champlain Valley. Chapman combined the marine levels as The Champlain Sea. He defined an upper, Coveville Stage and a lower, Fort Ann Stage comprising Lake Vermont. Stewart (1961) added an even higher Quaker Springs Stage. LaFleur (1965) working in the Hudson Valley, suggested that the lowest levels of Lake Albany in the Hudson Valley were coextensive with the upper levels of Lake Vermont in the Champlain Valley. Connally (1968) working in the uplands between the Hudson and Champlain Valleys confirmed LaFleur's suggestion. Connally and Sirkin (1971) altered existing terminology by restricting the name Lake Albany to the highest lake in the Hudson Valley, dropping the provincial name Lake Vermont, extending the names Lake Quaker Springs and Lake Coveville to the coextensive lakes, and using the name Lake Fort Ann for the lowest freshwater lake in the Champlain Valley. Lakes Albany, Quaker Springs, and probably Coveville extended all the way south to the Harbor Hill Moraine across Staten Island, New York.

The Woodfordian glacier, in its northward recession up the Hudson Valley, fronted an expanding Lake Albany. The Luzerne readvance took place during the existence of this lake and presumably the deposition of the Burlington till. With retreat of the Burlington ice margin into the Champlain Valley, the land to the south rebounded differentially causing a relative lowering of the lake level and formation of Lake Quaker Springs. Stewart and MacClintock (1969, 1970) projected Lake Quaker Springs northward to the Lamoille River, 15 miles north of Burlington. They state (1969, p. 163) that in this general area "the shore line features are so well developed that they seem to indicate that the Quaker Springs Lake was in existence for an interval as long as the later lake stages". However, good evidence for this lake is lacking north of Brandon in the Brandon and Middlebury quadrangles and Connally and Sirkin (1972) and Connally (1972) place the ice margin in the vicinity of Brandon, and Ticonderoga, New York, during Lake Quaker Springs.

As the ice margin retreated northward the land continued to rebound and the outlet shared by Lakes Albany and Quaker Springs was evidently breached forming Lake Coveville at a lower elevation. The ice retreated to near Burlington, readvanced to Bridport, and then retreated at least as far as the Lamoille Valley; all during the existence of Lake Coveville. Finally, the ice retreated to the north end of the Champlain Valley, the land continued to rebound, and a probable dam near Fort Ann, New York, formed Lake Fort Ann. Lake Fort Ann was most likely dammed to the north by the glacier as it stood at the Highland Front Moraine about 12,600 yrs. B.P.

Retreat of the Burlington ice north of the St. Lawrence Valley allowed Lake Fort Ann to drain northward down to lower levels (see Wagner, 1969). Following a short erosional interval (Stewart and MacClintock, 1969, p. 178) the Champlain Valley was invaded by marine waters to form the Champlain Sea. Coldwater marine molluscs in clays and sands between Vergennes and Burlington document at least one stage of the Champlain Sea in the field trip area.

FIELD TRIP STOP DESCRIPTIONS

Topographic 15 minute quadrangles covered: Burlington
Middlebury
Ticonderoga
Brandon

STOP 1. SHELBURNE VILLAGE SECTION: This is the type locality for the Shelburne drift described by Stewart (1961, p. 102) as "a small stream valley, one and one-quarter miles south-southwest of Shelburne Village. The Valley walls ... expose a layer of dark gray till over bedrock ... overlain by fifteen feet of red-brown sandy till that is covered by four to eight feet of bouldery lacustrine clay. ... The orientation of pebbles in the gray till show a fabric with maximum approximately north 30° east. The fabric of the overlying till is north 15° west." The lower till was later named the Shelburne till. Many workers have subsequently visited this section; some have supported the two-till interpretation while others have challenged it.

STOP 2. LEWIS CREEK DELTA: A gully exposure in this marine delta, 1500' south of Lewis Creek off Rt. 7 displayed the following section in 1965:

2' Sand, pebbly, probably marine; at 200' elevation.

15' Clay, gray, with scattered shells of marine clams.

8' Clay, brown, bouldery, probably lacustrine.

5' Till, gray lodgement; boulder pavement at top.

1' Sand, brown, stratified.

A very well formed beach ridge nearby at 250' may mark the high stage of the Champlain Sea or a post Lake Fort Ann stage, called "Lake New York" by Wagner (1969).

STOP 3. LITTLE OTTER CREEK SECTION: The composite section along the creek two miles north of New Haven shows the following:

2-10' Clay, bouldery, with stratified lenses of silt and sand; lateral gradation to till.

3-10' Till, clay-rich with boulders of varved clay; fabrics are N 4°W, N 6°W, and N 35°W.

3' Varved clay in situ.

14' Till, brown, bouldery, lodgement; fabrics are N 35° E and N 19° E.

15' Sand, pebbly, poorly stratified and interbedded till.

The upper till may be assigned to the Burlington Stade, the lower to the Shelburne (Calkin, 1965). Interbedded tills and lacustrine deposits suggest an active oscillating ice margin.

STOP 4. BRISTOL KAME TERRACE - DELTA: The ice contact deposits as first described by Chapman (1942) appear to be topped by a deltaic surface (village and airport) of Lake Coveville at 570'. Wave erosion at this level may have carried gravel out over the ice contact gravels to form the foreset-like beds seen at the outer edge (Stewart and MacClintock, 1969). Weak bars at the Lake Fort Ann level (420' here) occur nearby.

STOP 5. THE COBBLE AND KETTLED KAME TERRACE: Five miles south of Bristol off Rt. 116 is the Cobble, a bedrock outliner which has controlled the great width of the kame terrace here. Two kettle holes over 40' deep in the surface at 540-580' are below the level projected for Lake Quaker Springs. Do these kettles preclude the existence of Lake Quaker Springs here ?

STOP 6. CHIPMAN HILL, MIDDLEBURY AND LUNCH. This hill has more than 400' of relief, has exposures of bedrock near the base at the north, but only till is found at the surface within the upper 300'. Is it a drumlin ?

STOP 7. WEST BRIDPORT SECTION: This section was described by Connally (1970, p. 11). In 1964 the exposure showed:

0-2' Silty-clay containing ice-rafted(?) pebbles and boulders.

16' Silt and sand, laminated to thin-bedded, lacustrine.

5 1/2' Clay-loam till, dark gray (N3), with a lower 12-18" gray-black (N2) till overlain by 12-18" of oxidized gravel at the base.

3' Sandy-loam till, light olive-gray (5y 5/2), calcareous sandy-loam till.

Bedrock with striae oriented N 10° E.

Till fabric maxima are N 50° E for the olive-gray till and N 30° W for the overlying dark gray till. This agrees with the definition of NE Shelburne overlain by NW Burlington as seen at Stop 1. However, Connally (1970, Table 2) attributes the striae and both tills to the Burlington advance. The upper, bouldery, silty-clay is inferred to represent the Bridport readvance.

STOP 8. BRANDON-FORESTDALE DELTA: This delta was deposited by the Neshobe River. Chapman (1937, p. 59) inferred the Coveville level at 430' at Brandon but Connally (1970, p. 21) placed it at 405' farther south where Chapman's projection is 420'. Chapman attributed higher levels to local lakes but Connally correlated the well developed 500' level with Lake Quaker Springs. If time permits the 405', 500', and a higher 565' level related to the Lake Dunmore kame moraine will be visited.

STOP 9. LAKE DUNMORE KAME MORAINE: The kame moraine is part of a full deglacial sequence that consists of kame terraces surrounding Lake Dunmore, the moraine, outwash at Forestdale, and the eastern channel of the Brandon-Forestdale delta that is graded to a local lake level at 565'. We will drive through this sequence and stop if time permits.

STOP 10. COVEVILLE BEACH: Reworking of a kame terrace belonging to a sequence higher and earlier than the Lake Dunmore moraine is present north of the Middlebury River. This kame terrace has been reworked to form a sandy apron, probably a beach, at the base of the terrace at 480'. This is only about 20' below Chapman's projection for Lake Coveville. Because there is no beach between the pre-Lake Quaker Springs kame terrace and the Coveville level beach, the northern boundary of Lake Quaker Springs is inferred to be south of the Middlebury River, near Brandon.

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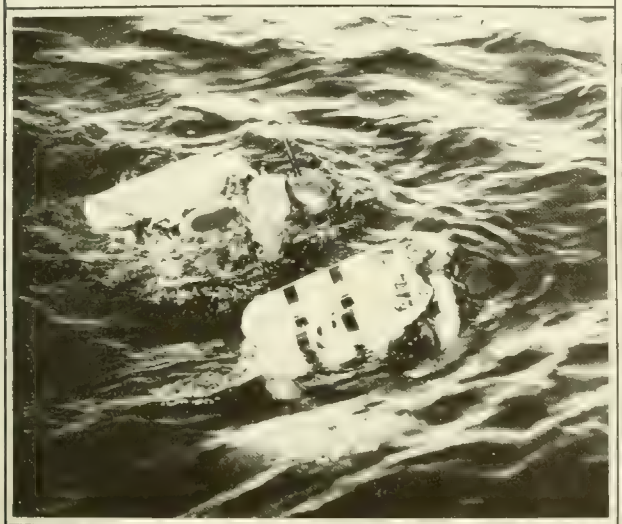
lake studies



Lake Studies Cover page: Various aspects of Lake Studies program, UVM Department of Geology. Photo at lower left by Robert Howe, UVM Geology Department. All others by Arthur Huse.

Trip LS-1

THE SLUDGE BED AT FORT TICONDEROGA, NEW YORK

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Introduction

While the legal battle has raged between Vermont and New York over the future of the sludge bed located at the mouth of Ticonderoga Creek, southern Lake Champlain, several comprehensive studies have been underway to determine the size, shape, and composition of the organic-rich deposit and its effect on lake geology, chemistry, and biology. Those involved represent Federal, State, and private organizations. Among them are the Federal Water Quality Administration, the U.S. Army Corps of Engineers, the University of Vermont and Middlebury College (FWPCA, 1968; FWQA, 1970, Folger, 1972).

Participants in this field trip will study the area (Fig. 1) from Middlebury College's research craft with such equipment as an echo sounder, surface drifters, Van Dorn bottles, secchi disc, dissolved oxygen kit, and grab samplers. They will thus gain a first hand look at the bathymetry, current regime, suspended matter and oxygen distribution and will be able to assess the effect of the pollutants on some physical properties of the sediments.

Creek Inflow and Sediment Characteristics

Ticonderoga Creek flows eastward from Lake George through the village of Ticonderoga to Lake Champlain over a distance of 5 km and vertical drop of 74 meters. Creek discharge at Lake

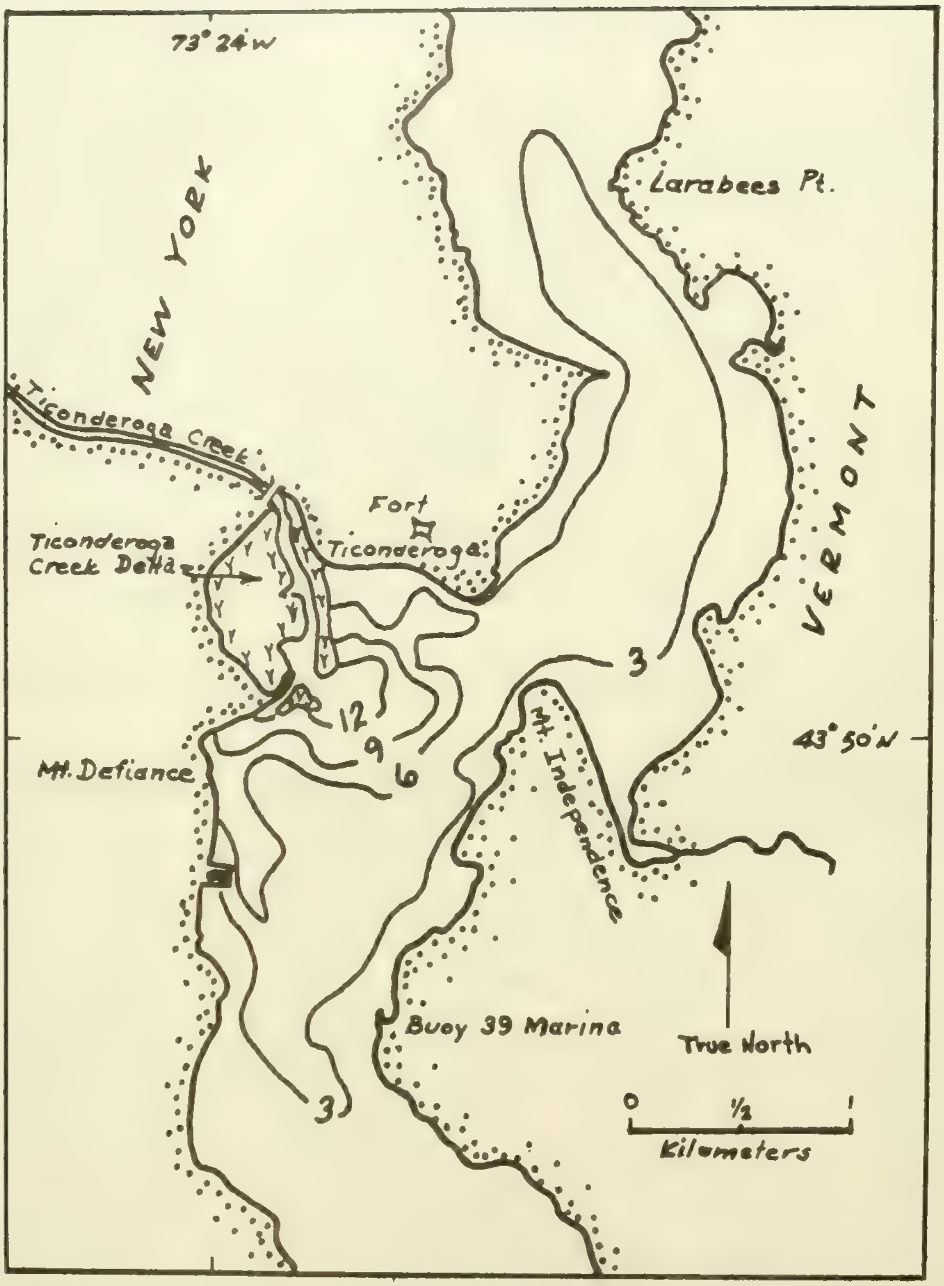

Figure 1. Contours showing organic carbon concentration (% dry weight) in bottom sediments of Lake Champlain.

George averaged 9 m^3/sec over a 17 year period with a maximum daily flow of 37 m^3/sec and minimum of about 0.2 m^3/sec (Wells, 1960). Because of five dams on the creek and because the creek and its main tributary, Trout Brook, both flow mostly over resistant metamorphic terrane where gradients are steepest the natural suspended load transported to Lake Champlain is probably low except during periods of high runoff. Below the dams, however, the creek carried abundant waste to the lake from the International Paper Company plant while it was in operation. Much of the delta built into the lake therefore consists of material derived from the paper plant and other industrial or municipal sources (FWQA, 1970).

The lake bottom drops off from the shallow delta front to the axis of the main lake channel where maximum depths are between 6 and 7 meters. The dark gray to black silty, sandy clays that cover the bottom near the creek mouth grade to very fine-grained greenish-gray clay (median diameter $<$ 2 microns) over a distance of several kilometers to the north and south. Acoustic penetration at 50 kHz of several meters in the finest textured clay is sharply reduced in sediments near the creek mouth. This is probably due to the coarser texture of the material but it may be due partly to the abundant gas in the most organic-rich sediments. Chlorite and illite are the dominant constituents of most bottom sediments in the southern part of the lake. Near Ticonderoga Creek, however, kaolinite becomes abundant with pollutants such as wood chips and fibers (Aubrey, 1971).

Pollutant Distribution

Because organic carbon is abundant in the paper plant waste, its distribution can be used to outline the sludge bed. Figure 1 shows contours of the organic carbon in bottom sediments. Highest values (∿14%) are concentrated near the creek mouth and decline to the north and south where few values exceed 2%. Some higher values have been measured in sub-bottom sediments (FWQA, 1970). Nitrogen is distributed in a similar pattern but concentrations grade only from about 0.4% to 0.2%. The highest values probably result mostly from raw sewage dumped into Ticonderoga Creek by the village of Ticonderoga (FWQA, 1970). Measurements of titanium, which is used as a whitener in the paper making process, have been made on sub-bottom samples collected on or near the delta. Concentrations range from 14.7% to less than 1% (FWQA, 1970). All three components decline in concentrations with distance from the creek mouth and appear to be good indices of effluent distribution on the bottom.

Flow Regime and Suspended Matter

The distribution of carbon also provides a rough guide to the flow regime of the lake in the area. Highest values, for example, extend farther north than south apparently as a result of the predominant northward flow of the lake. The smaller tongue of high values that extends southward probably is caused by the physiography of the delta which directs some creek flow southward especially on the west side of the lake and by reversals of lake flow when strong winds periodically blow from the north. Measurements

of surface currents in the fall of 1971 verified motion in both directions. Four surface drifters released in mid-channel north of Buoy 39 Marina moved northward at velocities between 6 and 12 cm/sec; two others released off the Marina moved southward at velocities between 1 and 5 cm/sec. Flow over the sludge bed is apparently sufficient to prevent oxygen depletion in bottom waters.

The concentration of suspended matter in bottom waters during the fall of 1970 ranged from about 15 to 20 mg/liter. Because higher values ($\sim$25 mg/liter) were observed north of the area shown in Fig. 1, it is doubtful that pollutants from the creek are primarily responsible for the high turbidity. Rather, most suspended matter probably consists of clay minerals stirred up by waves from the broad shallow shelf that surrounds most of the southern part of the lake.

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Trips LS-2, LS-3

SEDIMENTOLOGICAL AND LIMNOLOGICAL STUDIES OF LAKE CHAMPLAIN

by

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INTRODUCTION

Lake Champlain, one of the largest lakes in the United States, represents a major water resource for the northeast as well as a source of recreation, transportation, and municipal water. Before 1965, very few data were available on the lake. In that year, a cooperative study was undertaken by workers in several departments at the University of Vermont including biochemistry, botany, engineering, geology, and zoology, to gain a better understanding of the lake's past history, present condition, and future prospects. The purpose of this trip is to demonstrate the type of research being done, and report some of the findings.

We would like to thank Drs. Milton Potash and Philip W. Cook for contributing data on general limnology and phytoplankton, and Richard Furbush, Master of the UVM Melosira, for many successful cruises. Without the help of our graduate students, who have been credited where appropriate, this report would not have been possible. The work upon which this research was based was supported in part by funds provided by the U. S. Department of Interior as authorized under the Water Resources Research Act of 1964, Public Law 88-379.

Present Lake Champlain

Lake Champlain is approximately 110 miles long and has a maximum width of twelve miles, measured from the Little Ausable River, New York, to the shore of Malletts Bay, Vermont. It has a mean elevation of 92.5 feet above sea level and a water surface of 437 square miles (gross area 490 square miles). As discussed elsewhere (Hunt, Boardman, and Stein, 1971) Lake Champlain is composed of two morphologically distinct although interconnected north-south trending water bodies. The larger body is referred to as the main lake. The smaller water mass to the east, called the east limb, is connected with the main lake by three narrow passages. The lower third of the lake resembles a river in that its maximum width is one mile and its maximum depth 20 feet. The south end of the lake is connected with the Hudson River via locks of the Champlain Barge Canal. North of Crown Point, New York, the basin widens and deepens reaching a maximum depth of 400 feet near Split Rock Point.

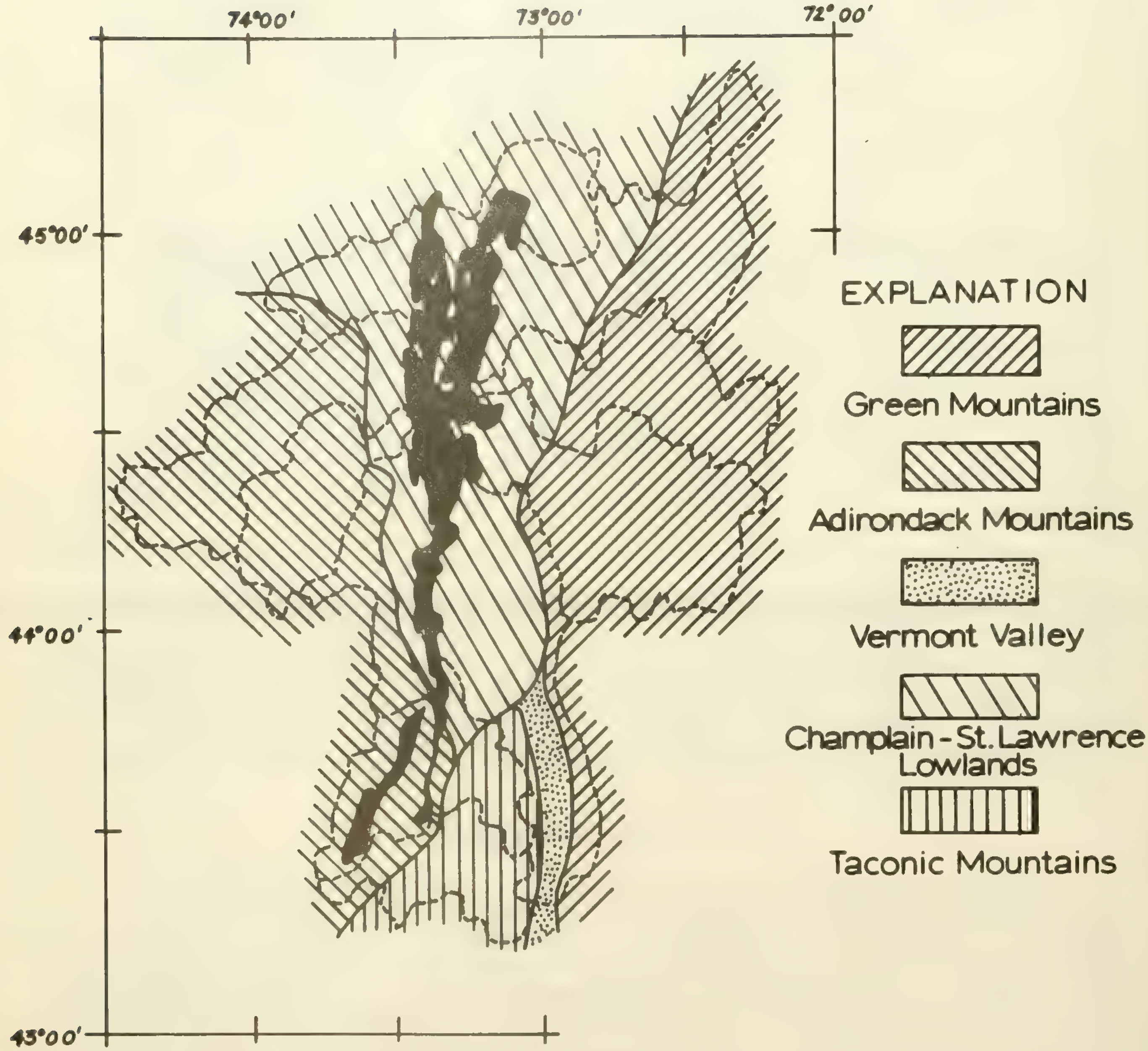
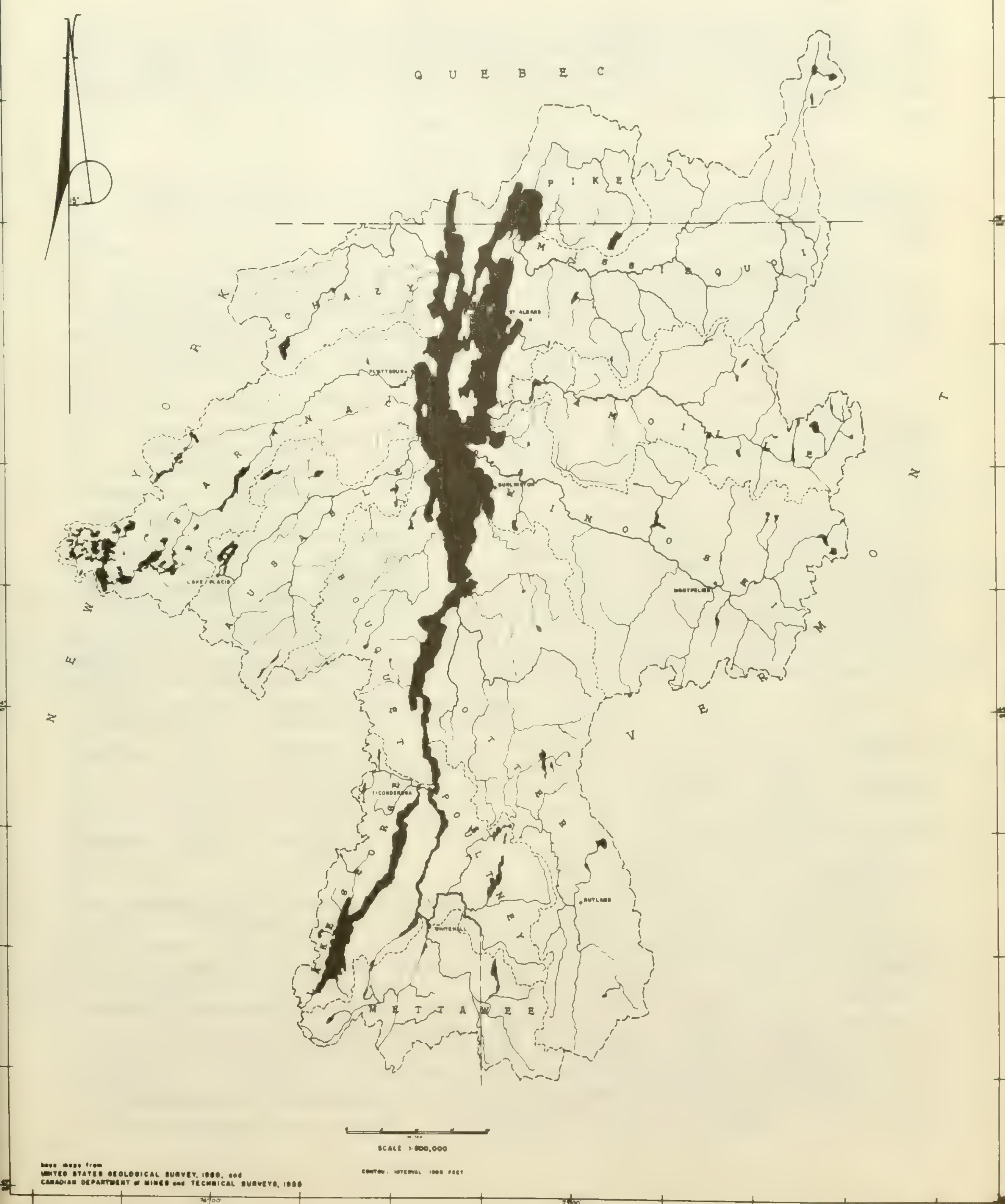


Figure 1. Morphological Regions of the Champlain Drainage Basin. The dashed lines designate drainage sub-basins. (From Hunt, Townsend, and Boardman, 1968).

Figure 2. The Lake Champlain Drainage Basin. (From Hunt, Townsend, and Boardman, 1968).



From Split Rock Point northward the lake becomes broader reaching its maximum width north of Burlington. Still further north it again takes on the character of a river, as it flows northward over a bedrock sill through the Richelieu to the St. Lawrence River. The mean discharge at Chambly, Quebec, is 10,900 cfs.

The drainage basin can be divided into five morphological regions (Figure 1). These include the Green Mountains, Adirondack Mountains, Vermont Valley, Champlain-St. Lawrence Lowlands, and the Taconic Mountains.

Fifteen hydrologic regions of the drainage basin have been delineated such that all of the tributaries in any region drain into a specified portion of the lake (Figure 2). This allows for an interrelated study between the drainage areas and the mineral budgets of the lake, a long-term study that is now in progress.

An inventory has been made of all of the streams that drain directly into the lake. Of a total of 296, only 34 have watersheds in excess of 10 square miles. In fact, 10 streams drain 80% of the total Champlain basin, and the 24 largest streams drain 95% of the total basin. There are 12 streams with drainage basins larger than 100 sq. miles; the largest is the Winooski watershed (1,092 sq. mi.). These streams discharge an average of approximately 12,000 cfs. of water into the lake. The volume of the lake is 912 x 10^9 cubic feet (Hunt, Boardman, and Stein, 1971), resulting in an average refilling rate of approximately 3 years. Some additional hydrological information has been summarized in figure 3.

	West Side	East Side	Total Basin
Catchment area:	2,618	5,126	7,744 sq.mi.
Percent of total area:	33.8	66.2	100
Mean discharge/sq.mile:	1.327	1.639	1.523 cfs/sq. mi.
Calculated total discharge into lake:	3,474	8,402	11,876 cfs
Percent of discharge into lake:	29	71	100

Figure 3. Provisional Summarized Hydrological Data for Lake Champlain.

The bedrock geology surrounding the lake basin consists of a diverse assemblage of rocks (Figure 4). High grade metamorphic

rocks of the Adirondack Mountains, mantled by unmetamorphosed sandstones and carbonate rocks, border the western margin of the lake. Unmetamorphosed or low grade metamorphosed carbonates, sandstones, and shales border the eastern margin and presumably underlie a large portion of the lake proper.

GEOLOGICAL HISTORY

The recorded geologic history of the Champlain basin started in the lower Paleozoic when sediments were deposited in marine waters that invaded eastern North America. These sedimentary rocks, which consist of limestones, shales, and sandstones, form the present lake basin. Thrusting from the east during the Paleozoic brought more highly metamorphosed rocks, which define the eastern margin of the lake basin, into contact with the relatively undisturbed basin rocks. The elongate shape of the basin, as well as the rapid change in the bedrock lithology across the lake, suggest that faulting may have played a part in basin deepening.

The history of the lake from the Paleozoic to the Late Pleistocene is not known, although for at least part of this interval the basin may have served as a river valley.

Evidence for glacial scouring is found today in the lake's ungraded longitudinal profile and in basins more than 300 feet beneath sea level. Presumably several times during the Pleistocene, ice occupied the lake basin. To date, however, no Pleistocene deposits older than Wisconsinan have been identified. Interpretations of the lake's Pleistocene history have been based primarily upon the recognition of former lake levels, some of which today are several hundred feet above sea level. The elevated shorelines are identified by ancient beaches, wave-cut and wave-built terraces, spits, and deltas. Chapman (1937) made a classic study of the lake history and a résumé of his findings is given here. Modifications of Chapman's regional framework include studies by Stewart (1961), and Stewart and MacClintock (1969). Chapman recognized three water planes. Two of these end abruptly when traced northward through the Champlain Valley. The highest plane can be traced to Burlington where it is at an elevation of about 600 feet. The middle plane, which rests about 100 feet beneath the highest plane, may be traced to the International Boundary. These two higher planes presumably terminate because they formed in a water body which abutted against the retreating ice margin. The lake in which these upper planes formed has been given the name Lake Vermont (Woodworth, 1905). During the time when the highest plane was formed, Lake Vermont drained southward through an outlet channel at Coveville, New York. The middle water plane (Fort Ann stage) formed at a later time when a new, more northerly outlet of lower elevation developed near Fort Ann. After the ice lobe had retreated sufficiently, the water level in the Champlain

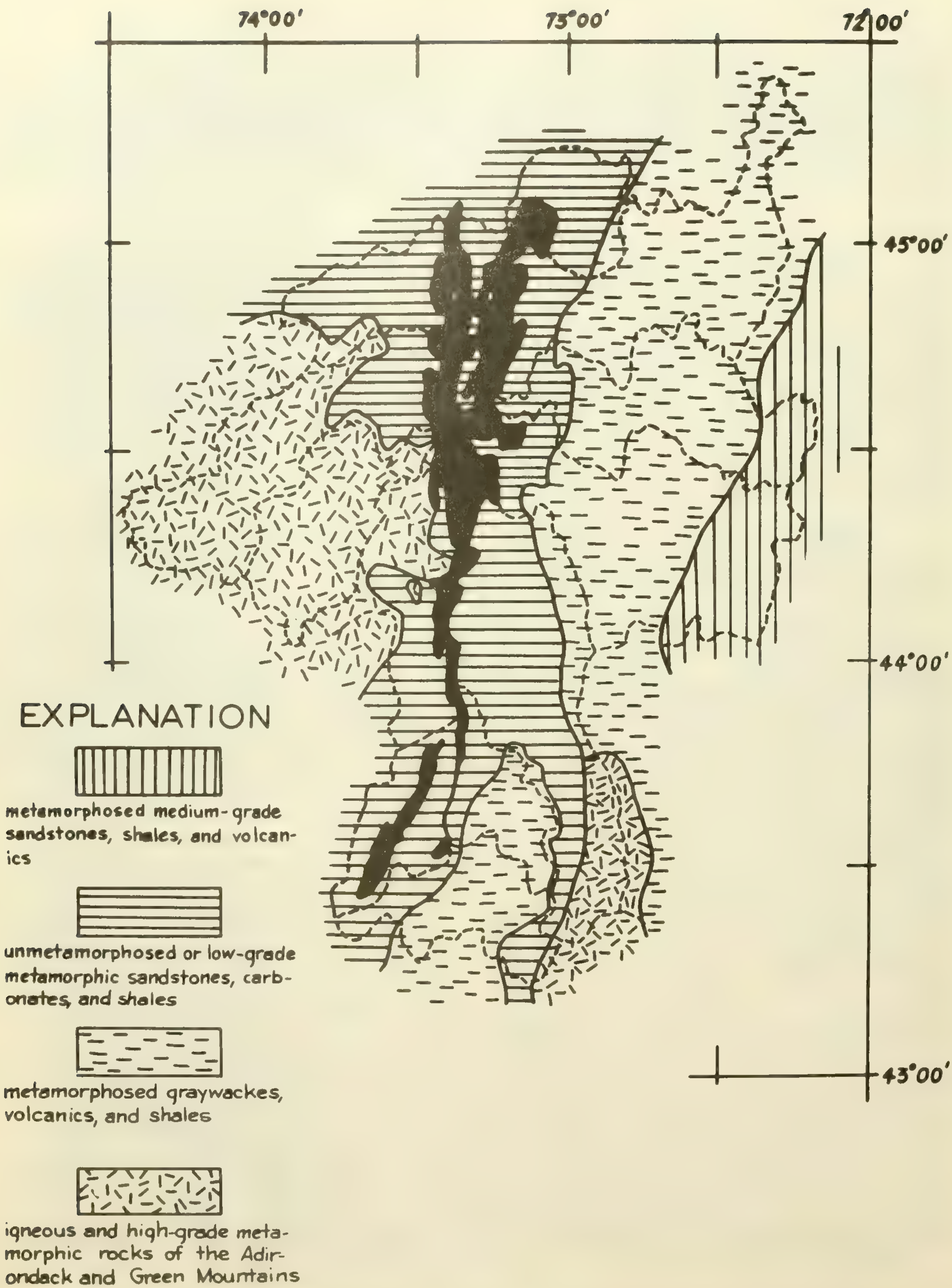


Figure 4. Major Rock Terrains of the Champlain Drainage Basins. The dashed lines designate drainage sub-basins. (From Hunt, Townsend, and Boardman, 1968).

Valley again dropped - this time several hundred feet, until it was continuous with marine water in the St. Lawrence lowlands. In the estuary which resulted, called the Champlain Sea, the lowest shorelines formed. Some time after the marine inundation the northern portion of the valley began to rise more rapidly than the southern portion. In time, the Richelieu threshold just north of the International Boundary was effective in preventing marine waters from entering the valley and the existing fresh water lake developed. Future tilting of only four-tenths of a foot per mile would cause Lake Champlain to again drain southward. This is only a small fraction of the tilting which has taken place since the Champlain basin was inundated by marine waters.

WATER PROPERTIES

Temperature:- The major portion of Lake Champlain can be considered to be a deep cold-water mesotrophic lake. Technically, it is classed as a dimictic lake (Hutchinson, 1957). This means that it has two periods during the year when the water in the lake is of equal temperature and is mixing. These periods of mixing alternate with periods of thermal stratification.

Thermal stratification begins to develop in June, and is well established in July and August. During mid-summer the metalimnion is at a depth of approximately 15 meters and includes the 12°C - 18°C isotherms. The period of summer stratification is short, for the depth of the thermocline increases steadily through August and September until the fall overturn takes place in October or November. Bottom temperatures in deep water remain at about 6°C during summer, but may rise to 12°C at the onset of the fall overturn.

The waters in the southern end and in the northeastern region of the lake are somewhat warmer than in the main lake, and warmer water is found in the bays along both shores.

During the winter most of the lake freezes over, and inverse thermal stratification develops with 4°C water at the bottom and 0°C water under the ice. Freezing begins in the narrow southern end, in the northern end, and in the northeastern portion of the lake. The wide main body of the lake is the last to freeze. In mild winters this portion may remain open throughout the winter season. The duration and intensity of the freeze depends on the severity of the winter.

Transparency:- The transparency of the lake, measured with a Secchi disc, ranges from about 3 to 6 meters. The deeper readings are encountered in late summer when algal growth is less. The disc reading in the southern part of the lake is usually less than 1 meter. Legge (1969) measured the penetration of light in

the lake, using a submarine photometer. Ten percent of incident light was usually found at a depth of 3 meters, 5 percent at 5 meters, and 1 percent at approximately 10 meters. The level of 1 percent incident light is therefore above the level of the thermocline.

pH and Alkalinity:- Champlain is an alkaline lake. The pH of surface water is above 8.0, but in the deep water the pH may get as low as 7.3.

The total alkalinity in the main lake, predominantly as bicarbonate, ranges between 38 and 46 mg/l, and averages 41 mg/l. Alkalinity values are higher in the southern end of the lake, and minimal values are found in the water in the northeastern sector. Abnormally high values are sometimes encountered at stations close to shore, modified by tributary inflow. The alkalinity at Rouses Point, near the outlet of the lake, is actually less than that of the main lake.

Major Cations:- The four major cations (Ca, Na, Mg, and K) have been measured in the lake and the results are summarized in Potash, Sundberg, and Henson (1969a). In the main lake the concentrations of these four cations are ranked in descending order as Ca, Na, Mg, and K, with median values of 15.8, 3.9, 3.6, and 1.1 mg/l. In the southern part the descending rank order is Ca, Mg, Na, and K, with median values of 24.4, 5.8, 5.1, and 1.2. In flowing from the south to the central lake, the water is diminished in the concentration of all four cations, especially in magnesium. The concentrations in the northeastern region of the lake are significantly less than in the main lake. In this part of the lake the descending rank order is Ca, Na, Mg, and K, the same as for the main lake, but the median values are 13.2, 3.0, 2.9, and 1.3 respectively. It is suspected that these differences between the main lake and the northeastern portion of the lake are influenced, in part, by ground-water intrusion, while the differences between the main lake and the southern lake are a result of surface inflow.

Major Anions:- The dominant anion in the lake water is the bicarbonate ion, which is mentioned under alkalinity. A few determinations have been made of the chloride and the sulphate ions. In the main lake the median concentration of sulphate is 15.4 mg/l, and of Cl, is 5.7 mg/l. The pattern for these anions is the same as for the cations; values are higher in the southern end of the lake, and lower in the northeastern part of the lake.

Dissolved Oxygen:- The concentration of oxygen dissolved in the lake water is one of the more significant parameters measured in lakes; it is essential for respiration for all animals and most plants, it facilitates the decomposition of organic matter in the lake, and it serves as an index for the general quality of the lake water. The major sources of oxygen dissolved in the

water are from exchange with the atmosphere and from photosynthesis by the plants in the lake. Oxygen is lost through respiration, decomposition, and increased temperature. The crucial test is the amount of oxygen in the deep water below the thermocline. In the deep water there is no source of new oxygen, and the supply that is there when stratification begins must last for the entire summer until the fall overturn mixes the water and carries down a new supply.

The trophic standard of a lake is sometimes measured by the concentration of dissolved oxygen in the deep water. In an oligotrophic lake the amount of organic material in the deep water during the period of summer stratification is of such small magnitude that oxygen consumed by decomposition has little effect on the concentration of oxygen in deep water. In a eutrophic lake, however, decomposition in deep water is great enough to reduce significantly the concentration of oxygen.

The main body of Lake Champlain is considered oligotrophic to mesotrophic by the oxygen standard. The lake water was more than 90 percent saturated in April, 1967, after the break-up of the ice cover. The oxygen in deep water from August through October was slightly less than 80 percent of saturation.

In some sheltered areas of the lake, for example, Malletts Bay, deep-water oxygen may be reduced to less than 1 percent of saturation (Potash, 1965; Potash and Henson, 1966; Potash, Sundberg, and Henson, 1969b). These are considered to be eutrophic areas of the lake.

BIOLOGICAL ASPECTS

Phytoplankton:- The phytoplankton is dominated by diatoms and blue-green algae. *Asterionella*, *Diatoma*, *Melosira*, and *Fragilaria* are dominant genera during the spring. *Ceratium* may become the dominant organism during mid-summer and the late summer-autumn plankton is characterized by the abundance of *Tabellaria*, *Gomphosphaeria*, and *Anabaena*. Overall, the phytoplankton is characteristic of a mesotrophic lake. Muenscher (1930) described the algae of the lake for 1928. Sherman (1972) has studied the diatoms in lake cores.

Zooplankton:- Ten species of Copepods (7 genera) and 12 species (9 genera) of Cladocora have been recorded from the lake. Among the Cladocera, *Bosmina*, *Daphnia*, and *Diaphanosoma* were the most abundant and widely distributed. *Diaptomus* and *Cyclops* were the only ubiquitous copepods. *Dinobryon* was found to be the most common Protozoa. Legge (1969) has described the seasonal distribution of the calanoid copepods in the lake.

Benthos:- The shallow-water (littoral) benthos consist of the usual communities of molluscs and insect larvae. The deep-

water fauna in organic silt consists of small worms, the glacial relic shrimp Pontoporeia, small clams, and a larval chironomidae.

Relic Pleistocene Fauna:- The fauna of Lake Champlain includes several species that are considered to be relics of the Pleistocene. Most of these animals are small invertebrates associated with the cold, deeper waters of the lake. They are mainly among the Crustacea. The schizopod species Mysis relicta (Opossum shrimp), a form common to the Atlantic Ocean, is found. Another inhabitant is the amphipod shrimp, Pontoporeia affinis, which was discovered in this lake only within the last five years. Both of these animals are common in the Great Lakes, but apparently are not very abundant in Lake Champlain. According to present thought these two species were able to inhabit the Pleistocene proglacial lakes and migrated from the Baltic Sea area during the Pleistocene, using a path around the Arctic Ocean, down through the Canadian chain of lakes, through the Great Lakes, to Lake Champlain (Ricker, 1959; Henson, 1966). Lake Champlain represents a terminus for these animals. Pontoporeia has not been found north of the St. Lawrence River east of the Ottawa River. Presumably an ice block prevented their migration into this area of the continent. There are some other animals in the lake which also are considered to be glacial relics. Among the small crustacean zooplankton would be included Senecella calanoides, which was first described from one of the Finger Lakes of New York, and Limnocalanus macrurus.

STRATIGRAPHY AND SEDIMENTARY HISTORY OF THE LAKE

Recent Sediments

The sediments exposed on the lake bottom today consist predominantly of materials deposited since the end of the Champlain Sea episode (about 10,000 years B.P.). The source of this material is (1) unconsolidated glacial deposits transported to the lake basin by streams (2) bedrock eroded from the shoreline; (3) organic matter from decomposing plants and animals; (4) biochemical constituents such as diatom frustrules. Based upon the size of past and present lake deltas, it is apparent that rivers have played an important role in transporting sediments to the basin. The present distribution of lake sands and gravels may be explained by wave winnowing. Coarse material, transported by streams to the lake is being left near shore. Fine material is being carried to the deeper basins. The lake muds, which constitute the deep water facies of the near shore sands and gravels, contain a significant fraction of organic matter, as well as biochemical constituents. With the possible exception of deltaic deposits which have not yet been studied, lake muds constitute the thickest sequences of recent sediments. Thicknesses of up to 80 feet have been observed (Chase, 1972). For purposes of discussion, recent

lake sediments have been grouped into four types - gravels, sands, lake muds, and iron manganese concretions. A description of these four sediment types is given below:

Gravels:- Gravel deposits, as defined by greater than 30 percent gravel (Folk, 1954), make up less than 4 percent of the sediments exposed on the lake bottom. Except for the gravel-sized material found in prerecent lake clays, gravels occur primarily in three areas: (1) shallow nearshore environments; (2) surrounding islands; and (3) at the mouths of rivers. Both in nearshore environments and surrounding islands the gravels represent lag deposits formed from the sorting of glacial till as well as the erosion of local bedrock. Gravels at the mouths of rivers are forming as a delta deposit. Former river channels can frequently be recognized by the distribution of nearshore gravel deposits.

Sands:- Sands, defined as having at least 30 percent sand (Folk, 1954), cover 22 percent of the lake bottom. The distribution of recent sand deposits is much like that of gravel in that they occur primarily in nearshore shallow water environments, and at the mouths of rivers (Fig. 5). In many areas they grade shoreward into gravels. The sands are low in organic matter, carbonate content, and are texturally and mineralogically immature.

Muds:- Muds cover approximately three quarters of the total area of the lake bottom. They occur primarily offshore in deep water (greater than 50 feet) where wave action is at a minimum, and in sheltered areas such as the bays. They are continuous through a facies change with recent sands. The surface of the muds is a grayish to reddish brown hydrosal. Beneath the interface, the muds are dark gray, uniform in grain size, and generally without lamination or structures although carbon smears and mottling do occur. The muds typically have a high organic content (up to 20 percent). The inorganic constituents of the muds consist of silica grains, clay minerals, and, in some areas, greater than 50 percent diatom frustrules.

Iron-Manganese Concretions:- "Manganese nodules" have been discovered in seven areas of Lake Champlain (Fig. 6). Only in the east limb, however, are they abundant and do they form well-developed concretionary structure. Here they occur in an almost pure state. In other areas the concretions are mixed with a terrigenous matrix which constitutes 90 percent or more of the sample. They occur primarily on shallow water platforms in water depths less than 40 feet and in areas where sedimentation rates are low (Fig. 6). Where concretions do occur at greater depths, they are found on slopes adjoining shallow water shelves, suggesting transportation off the shelves and down the slopes after formation. The nodules are associated with sandy sediments indicating that they are forming in high energy environments (Johnson and Hunt, 1972).

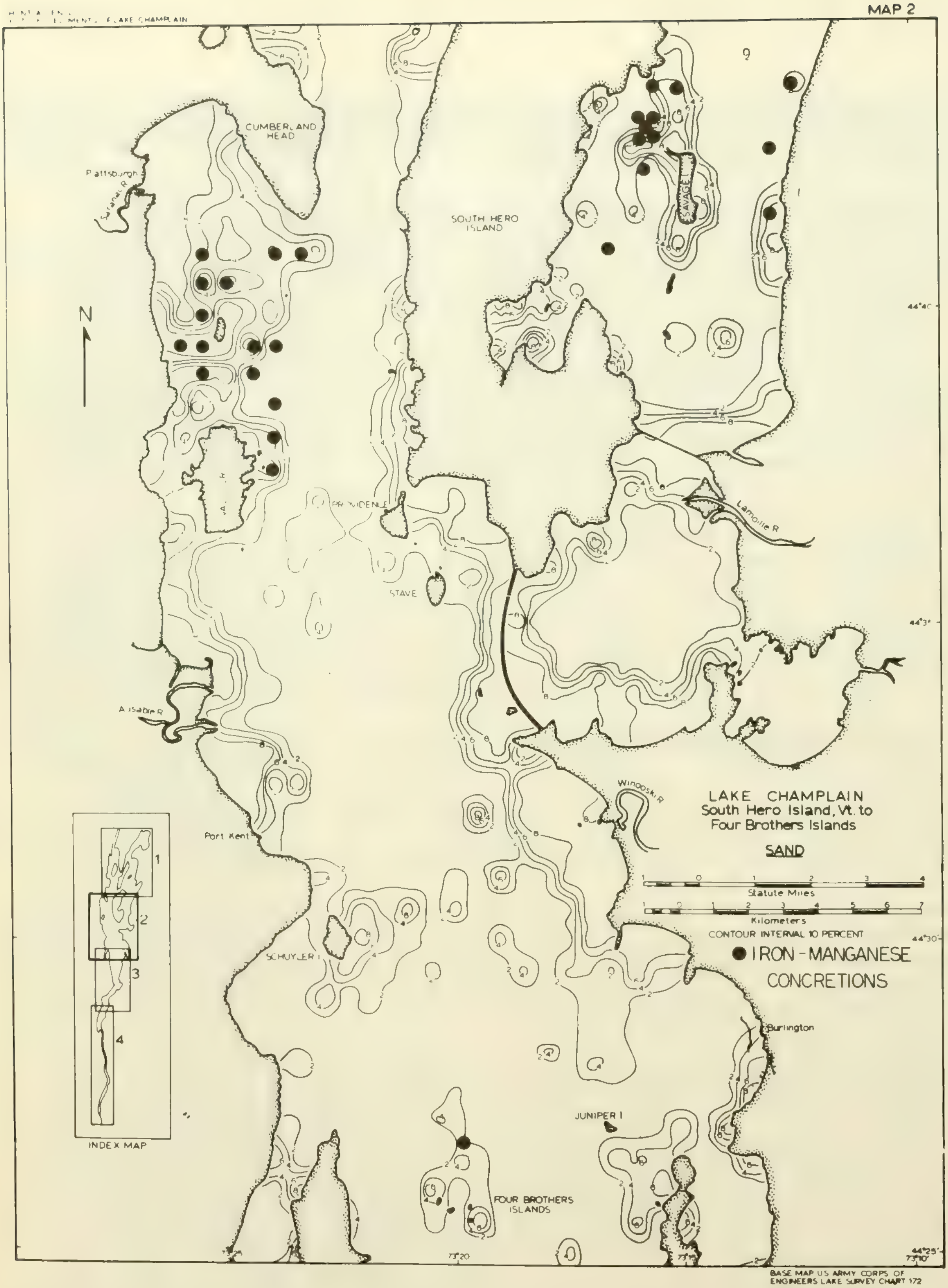


Figure 5. The Distribution of Sand in Central Lake Champlain.

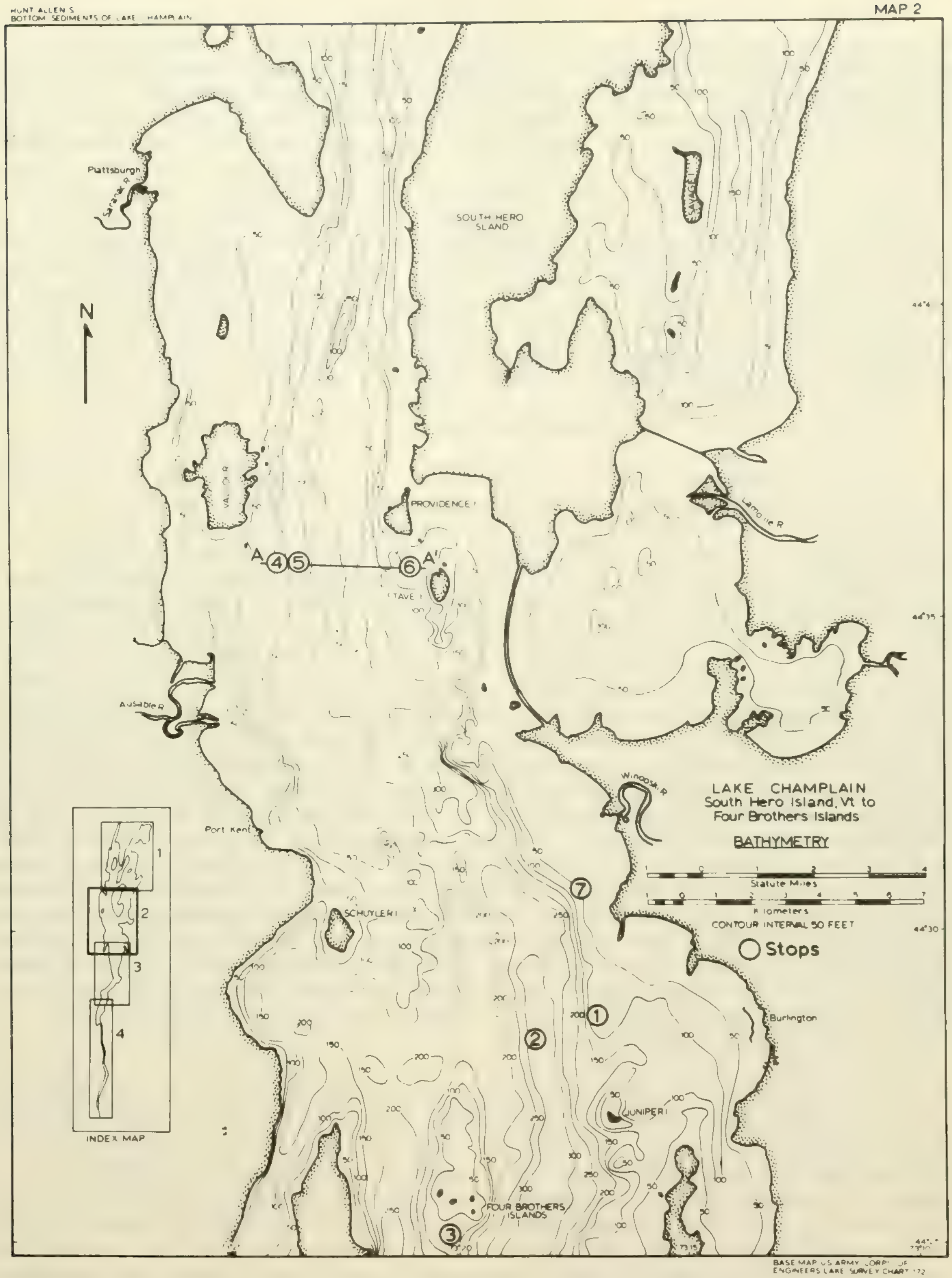


Figure 6. Bathymetry, Profile Locations, and Sampling Sites, Central Lake Champlain.

In size they vary from a few millimeters in diameter to greater than 10 centimeters; in shape, they range from spherical to reniform to discoidal. In well-developed concretions, light and dark brown bands may be seen in polished sections. Sand grains frequently form the nucleus. The geochemistry of the nodules was studied by Johnson (1969). The composition may be divided into two components, terrigenous and chemical. The terrigenous portion constitutes about one-third of the nodule by weight and consists predominantly of agglutinated fine-grained quartz and clay minerals. The dominant constituents in the chemical fraction are Fe_2O_3 38.5% and MnO 10.5% (east limb of lake). The nodules have a scavenging effect on trace metals. Maximum values of 410 ppm for cobalt, 605 ppm for copper and 585 ppm for zinc have been observed.

Prerecent Sediments

A fifth sediment type, lake clay, is exposed over a small area of the lake bottom, (1) on topographic highs where wave action is effective in erosion or preventing deposition of younger sediments and (2) where deep currents prevail. These sediments consist of two units. One was deposited in proglacial Lake Vermont during the retreat of the Pleistocene ice sheets, approximately 13,000 - 12,000 B.P. The second lake clay was deposited in the Champlain Sea which existed around 12,000 B.P. to 10,000 B.P. The maximum extent of both Lake Vermont and the Champlain Sea was considerably greater than present Lake Champlain. The only sediments of Lake Vermont or Champlain Sea origin which have been recognized in Lake Champlain are these lake clays. Presumably most of the coarser nearshore facies have been stranded away from the lake at higher elevations. These two units have been traced subsurface by sub-bottom profiling (Chase, 1972). Champlain Sea sediments, which have a known maximum thickness of about 100 feet, may be most easily recognized by the presence of foraminifera. More than a dozen genera have been recognized by Egolf (1972) from a single core at the mouth of Shelburne Bay. No fossils are known from the underlying Lake Vermont unit which has a total thickness in excess of 150 feet. Bedrock or till is believed to rest beneath these lake clays. A description of the lake clays, as exposed on the lake bottom today, is given below.

The clays are dense, sticky, and extremely poorly sorted, containing material ranging from clay size to cobbles several inches in diameter. Larger particles, including sand and gravel, are concentrated at the sediment-water interface and are dispersed through the sediment to a depth of at least several feet beneath the surface. The water-sediment interface is not a hydrosol as it is with the recent muds but is so well-compacted that it is difficult to penetrate. Wet, the clays typically are brown to yellow-brown in color at the interface and dark brown to dark grey below it. The clays also differ from the muds in their low organic con-

tent. The recent lake muds contain from 5% to 20% organic matter, whereas the lake clays have an organic content which rarely exceeds 5%. The lithology of the gravel fraction of the clay is variable and consists of metamorphic rocks, shales, and sandstones which outcrop within the drainage basin.

Sediments of Cultural Origin

In addition to the five naturally occurring sediments discussed above, sawdust, wood chips, paper waste, and cinders occur in several areas of the lake. Locally, as at the mouth of the Bouquet River, wood chips and sawdust, discharged during lumbering operations, and sludge at the mouth of Ticonderoga Creek discharged during paper production constitute a major portion of the sediment (See Folger, this guidebook). Cinders, which reflect the course of steamer traffic of past decades, are also abundant in some areas.

ADDITIONAL PROPERTIES OF SEDIMENTS

Chemical Properties:- The chemical properties of the lake sediments have not as yet been studied in any detail though several aspects are now being investigated. Notes from a few preliminary studies are, however, included here. Scattered carbonate analyses of lake sediments were made by Johnson (1967) who found carbonate percent to be remarkably low (less than 2 percent), considering the abundance of limestone bedrock in the drainage basin. The phosphorus content of the recent muds of St. Albans Bay was studied by Corliss and Hunt (1971) who found values twice as high in St. Albans Bay (1100 ppm) as in Lapan Bay, the control bay, which suggests nutrient build-up, presumably resulting from the discharge of sewage by the Town of St. Albans. Analyses of trace metals include the work of Cronin (1970) who analyzed the lead content in sediments taken on a traverse from Burlington Harbor westward to central Lake Champlain. He found that the highest concentrations occur in Burlington Harbor. Additional analyses by Hunt of lead, zinc, and chromium suggest that surface sediments have a higher concentration than underlying sediments, indicating cultural pollution. Chase (1972) analyzed for calcium and magnesium in sediment and interstitial waters of lake cores. Some variations were found but the interpretation of these data is not yet clear. An extensive study of trace metal concentrations utilizing core data is now underway by April (1972).

Organic Properties:- About 500 surface samples, primarily from central Lake Champlain, have been analyzed for total organic matter (Hunt, 1971). Values range from less than 1% to 22%. The data show that (1) organic content increases from the shoreline lakeward, (2) is positively correlated with increasing water depth, (3) increases with decreasing grain size. Numerous exceptions to these generalizations do occur. Some are easily explained, as in

the vicinity of Ticonderoga Creek where organic values of 22% were observed. This organic matter almost certainly is a product of industrial pollution (see Folger, this guidebook).

Mineralogical Properties

Heavy Minerals:- Townsend (1970) studied the heavy mineral content of sediments from the Ausable and Lamoille Rivers. He recognized two mineral assemblages: one, present in the western portion of the lake, is derived from the igneous and metamorphic rocks that occur in the drainage basin to the west. The second assemblage is best developed on the east shores of the lake and reflects the lower grade metamorphic source rocks which are exposed in the eastern portion of the drainage basin. In the central portion of the lake these two assemblages mix.

Clay Minerals:- Studies of clay-mineral distribution within the confines of Lake Champlain are in their infancy. Millett (1967) conducted X-ray diffraction studies of approximately 100 surficial sediment samples between Valcour Island, New York, and Thompson Point, Vermont. He found illite and chlorite are the dominant clay minerals with most samples having an illite:chlorite ratio in the range of 2.3 to 3.7. Some kaolinite is present in sediments from the Bouquet and Ausable Rivers which drain from the west side of the lake, but kaolinite appears to be essentially absent from lake sediments.

In addition, approximately 250 samples were checked by Bucke from 7 cores to determine clay mineral content vertically through the sediments. Millett's findings were supported in that the only significant clay minerals are illite and chlorite. The average illite:chlorite ratios in the cores range from 3.3 to 5.0 with a mean of approximately 4.2, about one higher than Millett's surface samples. This particular study was essentially a "shot-in-the-dark" with no previous knowledge of lithologies being penetrated by the cores. Current investigations of vertical clay mineral distribution is directed to determine if any consistent variations are detectable among Lake Vermont, Champlain Sea, and Lake Champlain sediments. Chase (1972) ran preliminary studies toward this end. Some suggestions of vertical variation are present, but as yet data is not consistent nor extensive enough to establish any real trends.

STOP DESCRIPTIONS

Stop 1. Telemetric buoy, Burlington Harbor. - We will pass alongside a buoy that has been installed by the Lake Champlain Studies Center to collect and transmit environmental data. The buoy normally transmits by radio every three hours to a base unit at the University which prints out the data in digital code. At present the buoy transmits information on wind speed and direction, air temperature, and water temperature at two depths. An accellerometer is being used in an attempt to measure sea state.

Stop 2. Reference Stations. - This stop, in 300 feet of water, is a reference station that has been sampled since 1965. A temperature profile will be obtained with a bathythermograph, demonstrating temperature variation with depth; and water samples will be collected at several depths for chemical analyses, and a plankton tow will be obtained for specimens of *Mysis*. In addition, a sample of recent lake muds will be collected.

Stop 3. Four Brothers Islands. - Cores from this area contain foraminifera indicating that these sediments were deposited in the Champlain Sea. The surface sediments on the Four Brothers rise contain a relatively high percentage of sand which distinguishes them from the encircling recent lake muds (Figure 5). The Four Brothers are situated on a topographic high and are subjected to extensive wave action, therefore, the sands may represent lag deposits rather than undisturbed pre-Lake Champlain sediments. The primary purpose of this stop is to obtain a piston core of Champlain Sea sediments.

Stop 4. Valcour Island. - This stop along with stops 5 and 6 will constitute a west-east traverse designed to show differences in thermal patterns, benthos, and sediment types across the lake. The interpretation of the sedimentary sequence based upon sub-bottom profiling data is given in figure 7. At this stop a bathythermogram will be taken and a grab sample will be collected and sieved for benthos. The waters surrounding Valcour Island were the site of the first naval engagement of the Revolutionary War.

Stop 5. Central Lake. - Sediments here are Champlain Sea deposits, underlain by Lake Vermont clays (Figure 7). At the surface the Champlain Sea deposits contain up to 40% sand as well as gravel-sized particles several inches in diameter, suggesting winnowing. Recent lake muds surrounding these Champlain Sea deposits indicate that restricted deep currents may be responsible for the winnowing.

Stop 6. Providence Island. - To complete the traverse profile a bathythermogram will be taken at this station, the benthos will be sampled, and a plankton haul will be made.

Stop 7. Winooski Delta (Optional) - This is a shallow-water stop at the mouth of the Winooski River. The influence of the river on the lake's bathymetry may be seen in figure 5. The sediment contains up to 80% sand. The area was considered by engineering firms as a source of fill in the construction of the Burlington beltline but a more economical land source was eventually selected. Deposition from the Winooski and Lamoille Rivers has virtually isolated Malletts Bay from the main lake. Note the tombolo forming the railroad crossing between the mainland and Grand Isle. As we return to Burlington Harbor, note the Champlain overthrust exposed on Lone Rock Point.

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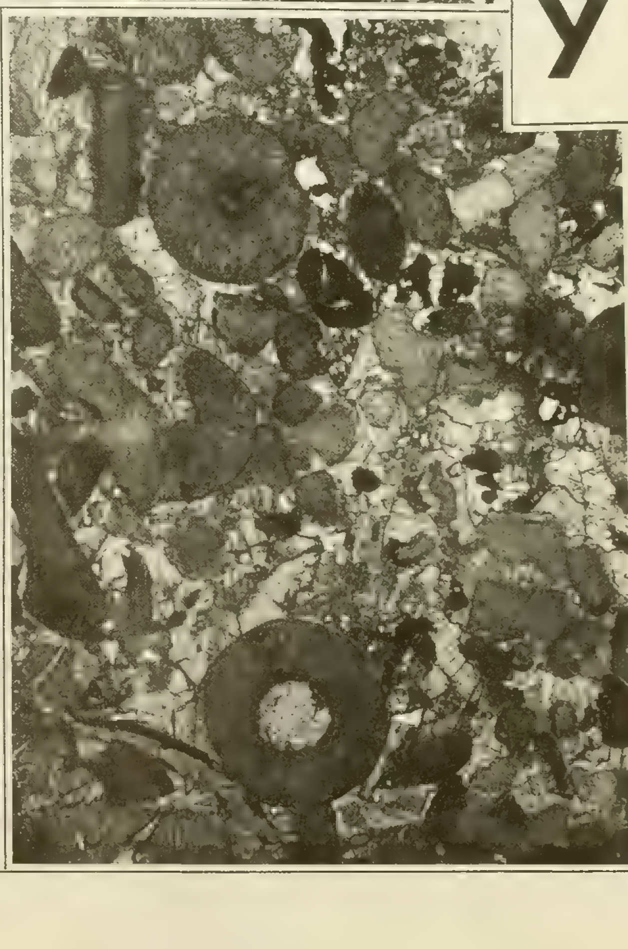
paleontology



Paleontology Cover page: Thin section photomicrograph of rocks from the Day Point Formation on Isle La Motte, western Vermont. For details see Finks, Shaw and Toomey (Trip P-1, this guidebook.)

Trip P-1

ORDOVICIAN PALEONTOLOGY AND STRATIGRAPHY OF THE CHAMPLAIN ISLANDS

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This trip focuses on the Chazy Limestone as exposed on southern Isle LaMotte, Vermont. The first part of the trip, conducted and described on the following pages by Shaw, covers the lower and middle parts of the Chazy at Stops 1, 2, and 3 on Figure 4. The second part of the trip, described by Finks and Toomey in the subsequent section, is a walking tour through the spectacular reef facies of the Chazy, starting 1/2 mile NE of Stop 2 on the same figure (Fig. 4, see also Fig. 2 of Finks and Toomey).

Paleontology and Stratigraphy of the Chazy Group (Middle Ordovician), Champlain Islands, Vermont.*

F.C. Shaw

Introduction

The Chazy Limestone (the oldest Middle Ordovician formation of the Champlain Valley) was first named by Emmons (1842) from exposures 15 miles north of Plattsburgh at Chazy, New York. Here and elsewhere in the northern Valley (Fig. 1) the unit outcrops on a variety of normal fault blocks. Given the low dips and heavy cover, Chazy stratigraphy is most easily understood from various shore outcrops around Lake Champlain. Valcour Island, SE of Plattsburgh, offers perhaps the best section of the Chazy and has been intensively studied (Raymond, 1905; Hudson, 1931; Oxley and Kay, 1959; Fisher, 1968; Shaw, 1968). The Isle LaMotte exposures to be covered on this trip are those studied by many of the same authors and, in addition, display the lower contact of the Chazy with the underlying Ordovician dolostones of Canadian age.

*Most of the following discussion and figures are excerpted from the trip run by Shaw for the New York State Geological Association (Plattsburgh, 1969) and from Shaw (1968).

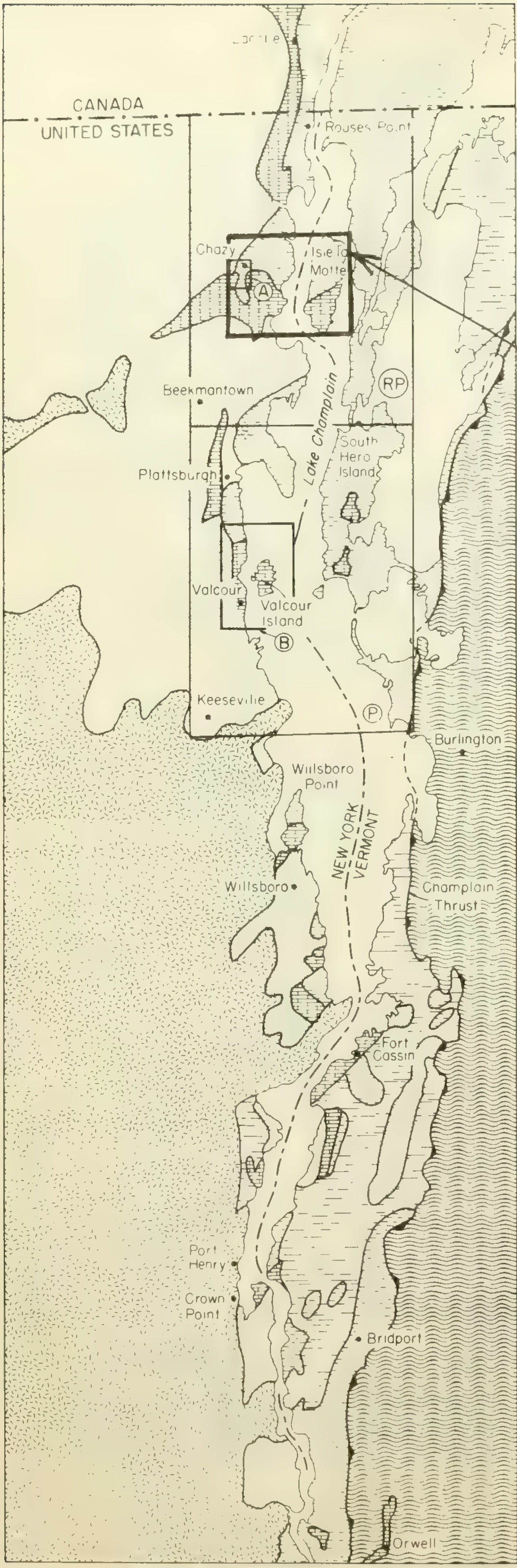

Figure 1. INDEX MAP
of
Champlain Valley
and portions of
New York, Vermont and Quebec
(area of figure 4)

EXPLANATION

Pleistocene (sands, clays)

Post-Chazy Ordovician sedimentary rocks
(mainly shales)

Chazy Group (limestones)

Pre-Chazy Cambrian-Ordovician sedimentary rocks
(sandstones, dolostones)

Cambrian-Ordovician metamorphic strata

Precambrian rocks
(gneiss, metanorthosite, charnockite, marble, quartzite)

Ⓐ Chazy area

Ⓑ Valcour area

Ⓟ Plattsburgh quadrangle

(RP) Rouses Point quadrangle

Geology modified from D.W. Fisher et al., (1962)

N

SCALE IN MILES

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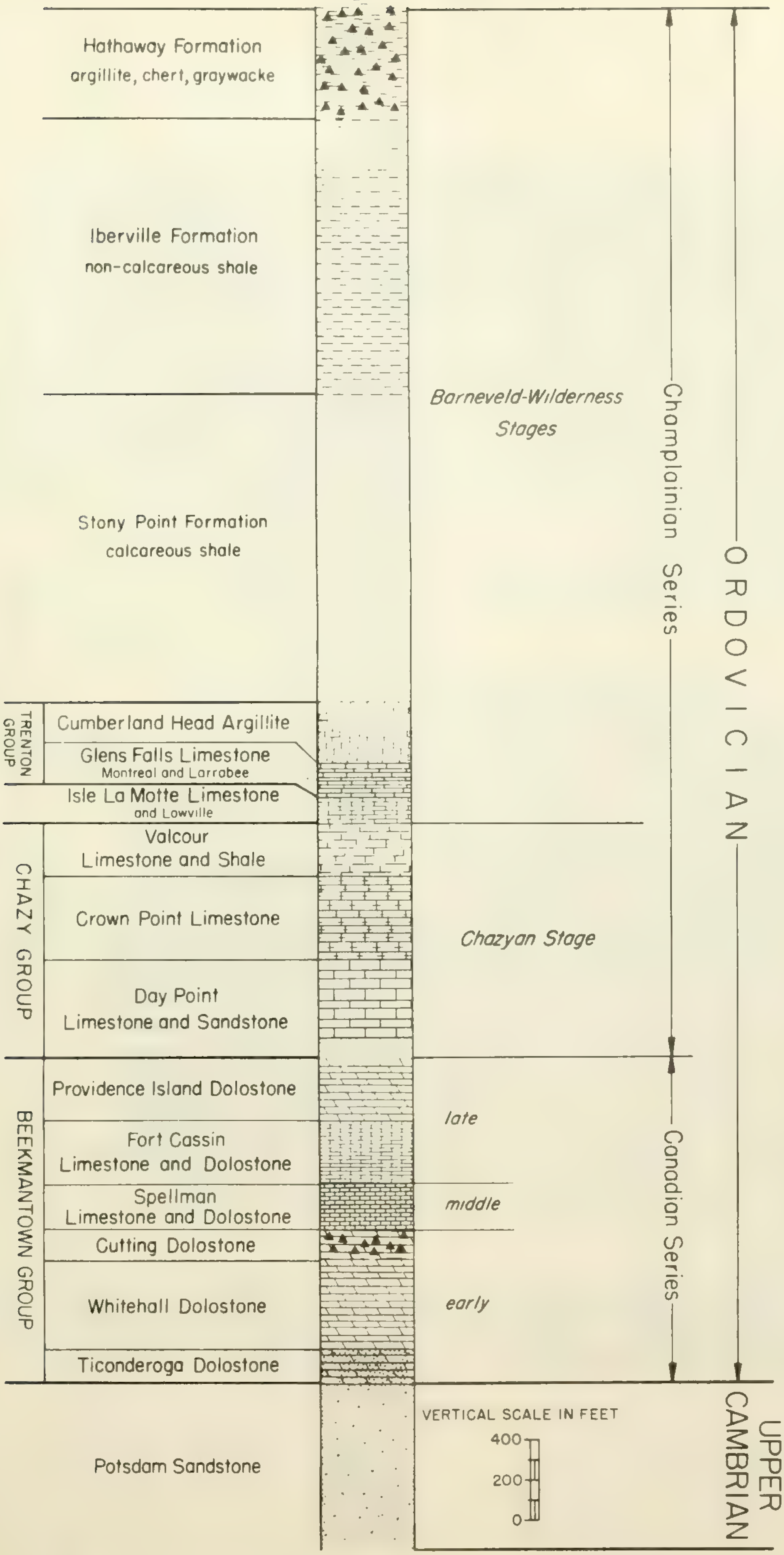

Figure 2. Generalized Stratigraphic Column—Champlain Valley

In the northern Champlain Valley (Valcour Island and north to the International Boundary), the Chazy Limestone (now group) consists of about 800 feet of quartz sandstones, calcarenites, dolomitic calcilutites and biohermal masses (Fig. 3). Three formations, Day Point, Crown Point, and Valcour, in ascending order, were proposed by Cushing (1905) and have persisted to the present, albeit with some controversy (Fisher, 1968; Shaw, 1968). Oxley and Kay (1959) further subdivided the Day Point and Valcour into members, those of the Day Point (Head, Scott, Wait, Fleury) coming from southern Isle LaMotte in the area to be visited. Shaw and Fisher experienced difficulty in using the Valcour subdivisions outside of their type areas at South Hero, Vermont.

DESCRIPTION AND INTERPRETATION OF CHAZY GROUP LITHOLOGIES

Day Point Formation

With the exception of the biohermal masses on Isle LaMotte the Day Point consists of a basal, cross-bedded quartz sandstone, followed by alternating units of shale, more sandstone, calcarenite, and topped with a relatively thick (35 feet) calcarenite unit (the Fleury Member). The lower sandstone, with its cross-bedding, presence of *Lingula* as nearly the only fossil, and overlying the supratidal Lower Ordovician, is probably transgressive and of very shallow water origin. This is further borne out by the presence of an oolite band in some of the sections around Chazy, New York. The source of the sand is unknown and no petrographic studies on this formation or on most of the other Chazy units have been undertaken. Derivation from Cambrian sandstones exposed on lowly emergent land to the west appears feasible. The calcarenites are primarily echinodermal in origin, although bryozoans and trilobites also occur abundantly, particularly in the Fleury (Ross, 1963, 1964; Shaw, 1968). Again, shallow water seems indicated, although probably subtidal judging from the abundant faunas (compare Laporte, 1968 and Textoris, 1968).

At Valcour Island and on the adjacent shore at Day Point, New York, the upper Fleury calcarenites are interbedded with dark, muddy limestones, some of which contain the varied silicified trilobite fauna described by Shaw (1968).

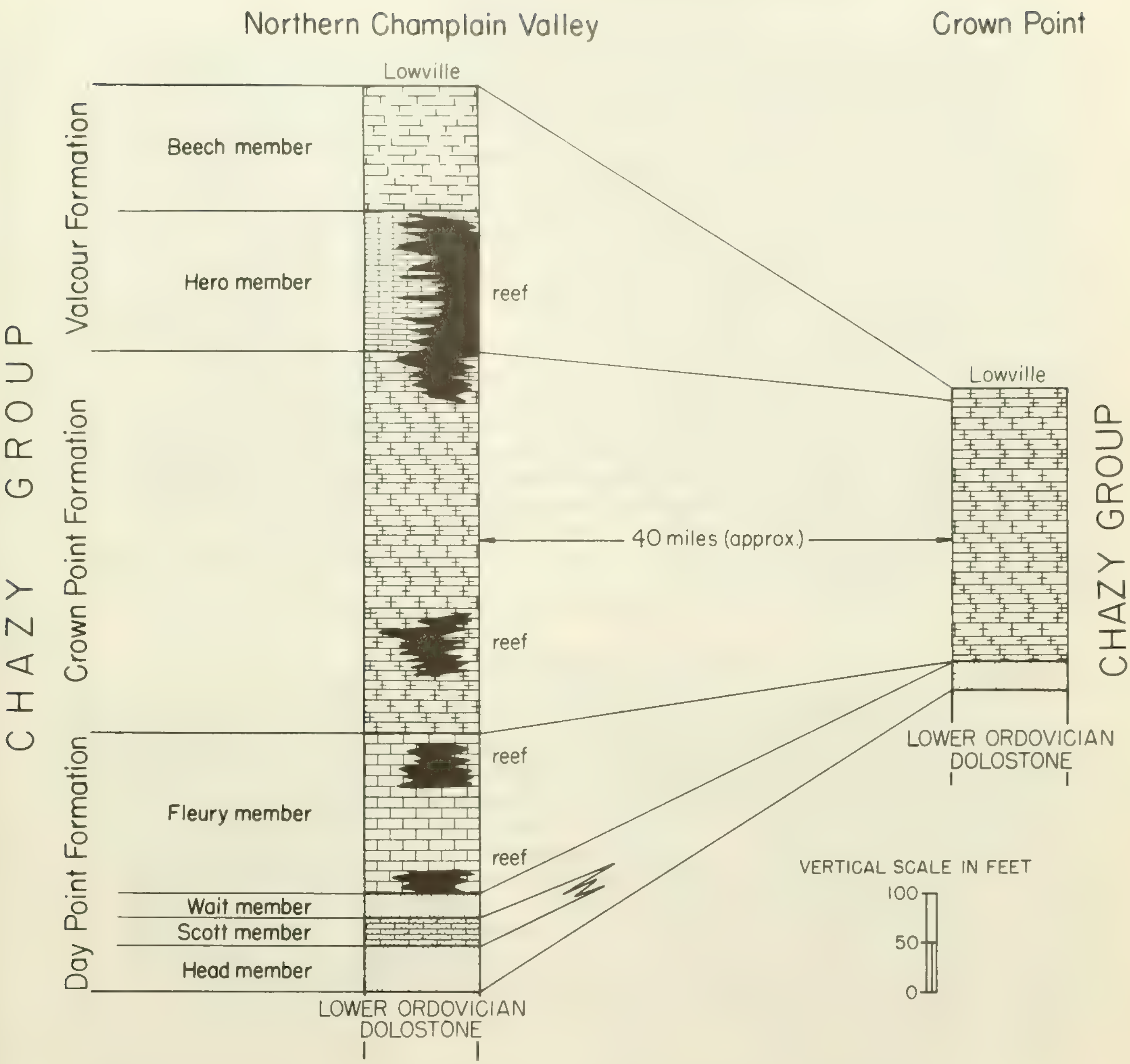

Figure 3. Lithologic Correlation

Crown Point Formation

The Crown Point Formation begins where muddy limestones become the dominant lithology. A striking feature of this formation is the abundance of thin (maximum 1/2 inch thick) dolomite stringers. Thin section analysis of many of these irregular stringers indicates that they are composed of argillaceous material, calcite grains, and scattered dolomite rhombs (Barnett, personal communication, 1969). Judging from the abundant faunas (gastropods, trilobites, ostracods, brachiopods) and their preservation (some trilobites and ostracods articulated), this lithology represents somewhat deeper and less agitated water. This leaves the origin of the dolomite to be explained inasmuch as recent discussions of dolomite have focused on a supratidal origin. Possibly this dolomite is secondary. Similar lithologies are known in the Ordovician of the southern Appalachians and Nevada and present a good petrologic problem. The 200-300 feet of the Crown Point Formation has never been subdivided into members, attesting to its homogeneity. Twenty-five miles south of Valcour Island, at Crown Point, New York, nearly the whole section (250 feet) is comprised of Crown Point Formation lithology (Fig. 3).

Valcour Formation

The Valcour Formation is characterized by a return to calcarenites, interspersed with limestones of Crown Point aspect. In addition, much of the Valcour as well as the underlying Crown Point displays well-developed bioherms consisting of stromatoporoids, bryozoans, calcareous algae, sponges and corals with an accompanying fauna of trilobites, brachiopods, cephalopods, and echinoderms. Spectacular examples of these will be covered on this trip. The channels in these reefs, the packing of these channels with trilobite and nautiloid fragments, and the accompanying carbonate sands again argue for relatively shallow water, with the more typical muddy limestones probably occupying slightly deeper basins between.

The Valcour is overlain by rock units usually assigned to the Black River Group, although outcrop or exact paleontological continuity with the type Black River of central New York and Ontario is difficult to demonstrate (Johnsen and Toung, 1960; Hofmann, 1963).

PALEOGEOGRAPHIC SETTING OF THE CHAZY GROUP

As mentioned above, the lower Chazy Group evidently represents a transgressive sequence over Lower Ordovician dolostones. The relative thinness, lack of abundant clastics, shallow water features, lack of volcanics and predominance of "shelly" rather than graptolitic faunas all argue for a setting on the platform or at best at the very edge of the miogeosyncline. Most paleogeographic reconstructions of Chazyan time (Kay, 1947) exhibit this relationship. Although the Chazy group thins and disappears southward and westward into New York State, lithologically and faunally similar units persist northward to the Montreal area and eastward into the Mingan Islands of eastern Quebec (Hofmann, 1963; Twenhofel, 1938; Shaw, ms.). Westward thrusting along Logan's Line has covered much of the miogeosyncline to the east, leaving us with either unfossiliferous or graptolite-bearing rocks which defy exact comparison to the Chazy Group. Speculation as to the exact geography of the Appalachian geosyncline in this area during Chazyan time is hazardous.

FAUNAS OF THE CHAZY GROUP

Raymond (1906) identified three faunal zones in the northern Champlain Valley Chazy Group, corresponding roughly to the three formations proposed by Cushing (1905). These were not really assemblage zones in the modern sense but relied heavily on two brachiopods and the large gastropod *Maclurites magnus* LeSueur (Pl. 1, Fig. 3). Raymond and later workers (Welby, 1961; Erwin, 1957; Oxley and Kay, 1959) also thought that the trilobite *Glaphurus pustulatus* (Walcott) first appeared at the base of the Valcour Formation. All of the above instances now appear to be examples of local abundance and/or facies control, although they are of some stratigraphic use locally in the Champlain area. Ross (1963, 1964), Cooper (1956), and Shaw (1968), using bryozoans, brachiopods and trilobites, respectively, were unable to make meaningful faunal subdivisions of the Chazy Group. Bergstrom (1971) has summarized the known information of Chazy conodonts. Nevertheless, the Group as a whole is distinctive, containing as it does the first appearance of stromatoporoids, primitive tetracorals, bryozoans (?), and primitive pelecypods. In addition, twenty-four genera of trilobites appear first in the North American Ordovician here in the Chazy Group. By contrast, graptolites and several long-ranging groups of trilobites such as robergioids and agnostids are absent from the Group, probably as a result of facies control or restricted oceanic circulation.

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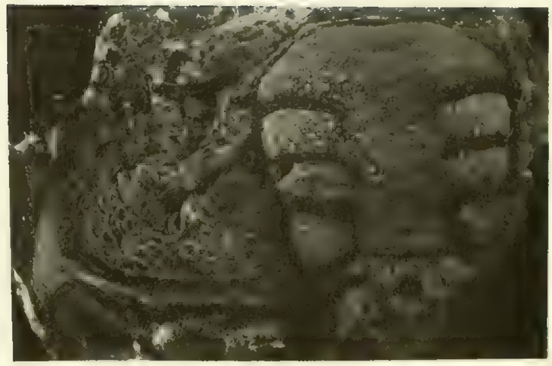

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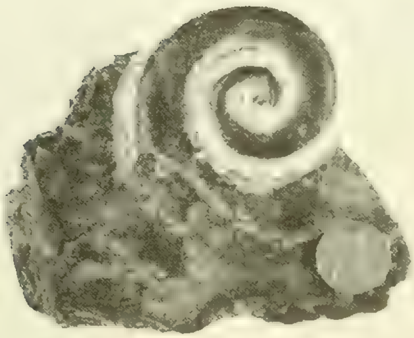

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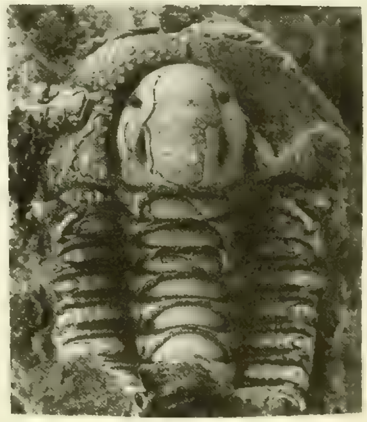

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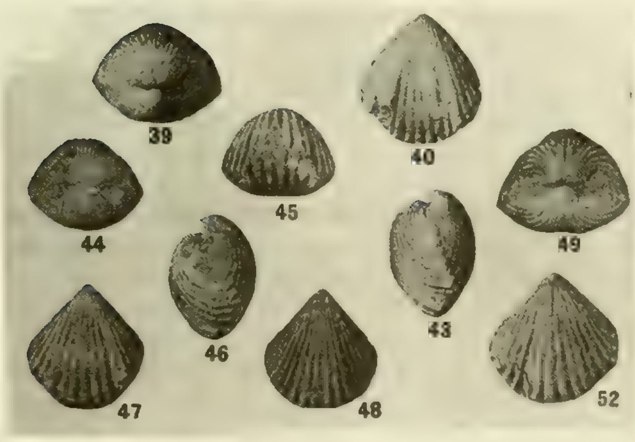

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Plate 1

1. *Amphilichas minganensis*: cranidium, dorsal view x2, from fine lime mud infilling reef framework at Sheldon Lane, New York.

2. *Paraceraurus ruedemanni*: cranidium, dorsal view x1, same lithology and locality.

3. *Maclurites magnus*: shell largely recrystallized. Crown Point Formation Intersection of NY 348 and I87, SW of Chazy Village.

4. *Glaphurus pustulatus*: partial cranidium and thorax, dorsal view x4. Same lithology and locality as 1 and 2.

5. *Rostricellula plena*: Valcour Formation, Chazy, New York. (From Cooper, 1956) x1.

6. *Pliomerops canadensis*: complete specimen, lower Valcour Formation, east side of Valcour Island, New York. x1.

Figures 1, 2, 4, 6 also appear in Shaw (1968)

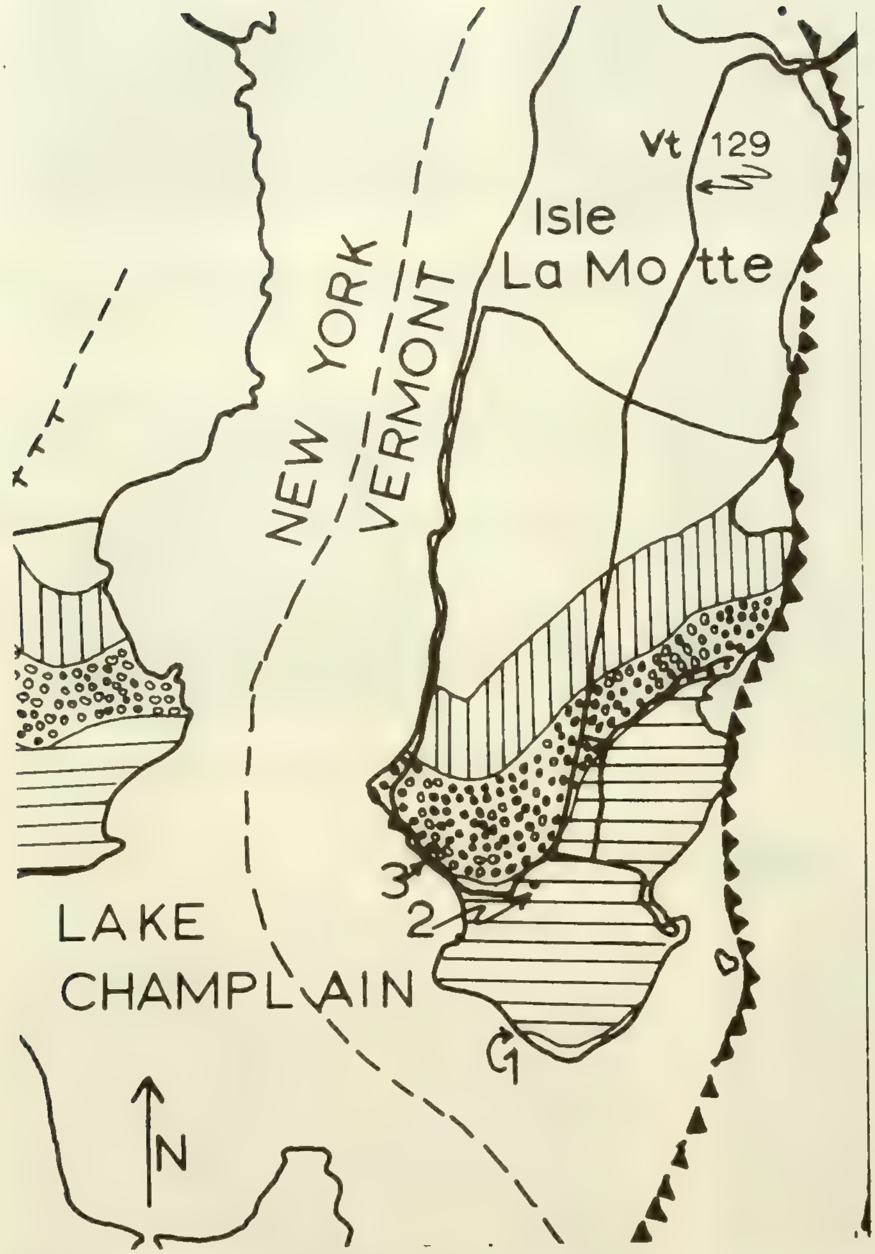

Figure 4. INDEX MAP OF THE ISLE LAMOTTE AREA, VERMONT

Horizontal Lines:	Day Point Formation
Dots:	Crown Point Formation
Vertical Lines:	Valcour Formation
Scale:	1 inch to 1 mile (Modified from Fisher, 1968)

In sum, the Chazy Group records a diverse marine fauna of cratonic aspect, including the very early representatives of a number of successful Paleozoic taxa. Exclusion of other taxa expected to be present, as well as facies dependence of organisms within the various facies of the Chazy Group (Shaw, 1968) generate some problems in correlating the Group to other North American Ordovician sequences.

FIELD TRIP STOPS

PLEASE NOTE: ALL STOPS ARE ON PRIVATE PROPERTY WHICH WE HAVE SPECIAL PERMISSION TO ENTER. DO NOT SMOKE IN THE FIELDS, KNOCK OVER FENCES, ETC. OTHERS WILL WANT TO RETURN TO THIS CLASSIC LOCALITY AFTER YOU.

Stop 1 - The Head, Isle LaMotte, Vermont, 3 miles SSW of Isle LaMotte Village. Lakeshore outcrops of the Providence Island Dolostone (Lower Ordovician) and the Day Point Formation (Chazy Group). Dip of both units several degrees to the north. Contact well-exposed and at least locally unconformable. Locality discussed by Shaw (1968), Erwin (1957). Section measured and described by Oxley and Kay (1959).

Approximately 40 feet of Providence Island Dolostone is exposed, being very fine-grained, massive, thinly laminated, and unfossiliferous. Mud-cracks and a few ripple marks complete the picture of a unit deposited in very shallow water. Following Laporte (1967) and Textoris (1968), the environments echo that of dolomite formation described from Florida and the Bahamas. No detailed petrologic work has been done on this unit. In the absence of fossils, the age of this unit is not known. The underlying Fort Cassin Limestone (not exposed here) is known to be Late Canadian.

The Chazy Group begins here with about 20 feet of quartz sand and siltstones together with minor amounts of greenish shale (Head Member of Oxley and Kay, 1959). Ripple marks and cross-bedding are common. The fossils consist primarily of "fucoids" (probably recording a variety of trails, worm tubes and the like) and *Lingula*. The succeeding Chazy unit (Scott Member of Oxley and Kay) consists of about 40 feet of echinodermal lime sand, cross-bedded in some places. Brachiopods (*Orthambonites?*) and indeterminable trilobite scraps are the chief recognizable fossils. The overlying 15 feet of quartz sandstone (Wait Member of Oxley and Kay) appears very similar to the initial sandstone.

This second sandstone is followed by a thick (115 feet) lime sand (Fleury Member of Oxley and Kay) which occupies most of Scott Point and The Head south of the road. Much of the unit is composed of echinoderm fragments, although little is known about the actual morphology of the creatures involved. Both the fragmental nature of the fossils and the frequently observed cross-bedding argue for considerable agitation of the ocean bottom.

Stop 2 - Same vehicle location as Stop 1. Upper Fleury Member of Day Point Formation and overlying Crown Point Formation. 200 yards south of the right angle bend in the road is locity R 25 (Shaw, 1968) which yielded 12 genera of trilobites, including Sphaeroxochus and Ceraurinella, from a particularly coarse pocket in the upper Fleury lime sands. Gastropods (Raphistoma) and brachiopods (Orthambonites?) are also present. This same stratigraphic level elsewhere, particularly 1 mile to the NE, displays spectacular bryozoan bioherms and the very early tabulate coral Lichenaria (Pitcher, 1964) to be seen later on this trip.

About 50 yards north of the road at this same stop, the silty, Maclurites-bearing limestones of the Crown Point appear. The actual contact with the Day Point is not visible but the lithologic change is evident. The Crown Point here contains several modest bioherms which have not been studied in detail. The earliest known stromatoporoids (Pitcher, 1964) are known to be important reef builders nearby in this unit and doubtless are dominant here as well.

Stop 3 - Fisk Quarry, 2.5 miles SSW of Isle LaMotte Village. Middle Crown Point Formation, consisting of fine-grained, dark, silty limestone with buff-colored dolomitic partings. This is "typical", non-reef Crown Point lithology. However, in the quarry wall and some of the cut blocks, small "reeflets" can be seen. These are assumed to be largely stromatoporoids and calcareous algae, although they have not been studied as intensively as the reefs at the same horizon to the east. Evidently, these reef masses could grow at some depth in relatively silty waters. The mechanism of their establishment thus does not appear to be tectonic. Maclurites (large gastropod, rare) and a few trilobites and brachiopods may possibly be collected from the limestone, although they are not abundant.

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PALEOECOLOGY OF CHAZY REEF-MOUNDS*

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REEFS AS BIOLOGIC COMMUNITIES

Geologists have commonly concerned themselves with reefs as a problem in the building and maintenance of a framework in the face of the destructive effects of wave-energy. From the point of view of the reef organisms the paramount aspect of the reef environment may well be the opportunities it provides for interactions between organisms. To use an analogy which may be appropriate on many levels, the former approach is like viewing a city as a problem in architecture, while the latter is like viewing a city in terms of social interactions. What draws people to cities is that they provide a maximum availability of functional relationships, or to use the equivalent biological term, of ecologic niches. The high population density of a city, as well as of a reef, is both the necessary condition for such functional diversity as well as an ultimate result of it.

As with cities, one of the important problems of a reef is the pollution of its environment by its own metabolism. In particular, the depletion of oxygen and the production of nitrogenous wastes are the most acute problems. The solution adopted by the inhabitants of modern coral reefs is the development of a symbiosis with certain phytomastigophora, termed zooxanthellae, which live in the tissues of reef organisms and have been shown to absorb nitrogenous materials. In living coral reefs zooxanthellae are present in scleractinians, sponges, and even the giant clam Tridacna. It is likely that the Scleractinia had this adaptation well back into the Mesozoic, for hermatypic Scleractinia and their reefs go back at least to the Jurassic, and they form the last part of an ecologic succession, or sere, in late Triassic reefs (Sieber, 1937). Whether this was true of Paleozoic reef organisms is not known. Certainly, calcareous algae were present in Paleozoic reefs and have continued to those of the present day. These free-living algae also contribute to oxygen replenishment and nitrogenous waste absorption. They must have been as essential to ancient reefs in this metabolic function as in their well-known frame-building properties.

TROPHIC RELATIONS IN THE CHAZY REEFS

The principal reef-building organisms of the Chazy reefs are stromatoporoids (Cystostroma, Pseudostylodictyon or Stromato-

* This discussion is a revised version of the trip run by Finks and Toomey for the New York State Geological Association (Plattsburgh, 1969).

cerium), lithistid sponges (Zittelella Anthaspidella), tabulate corals (Lamottia or Lichenaria, Billingsaria, Eofletcheria), bryozoa (Batostoma, Cheiloporella, Atactotoechus), and calcareous algae (Solenopora, Sphaerocodium or Rothpletzella, Girvanella). In addition, several trilobites (Glaphurus, Pliomerops and bumastids) and numerous pelmatozoan fragments are found in such intimate physical association with the reef as to be likely inhabitants of its surface. The large gastropod, Maclurites, and many genera of large nautiloids, are very abundant in the pelmatozoan-brachiopod calcarenites between the reefs, in calcarenite-filled channels cut into the reefs, and also in calcilutite-filled pockets or channels within the reefs. It is likely that these vagile organisms were also regular participants in the foodchain of the reef community.

Unlike modern reefs, in which macrophagous, carnivorous coelenterates are the dominant element, and feed upon an abundant fauna of small nekton, these Middle Ordovician reefs are dominated by suspension feeders. This is especially true when one considers the recent reinterpretation of stromatoporoids as sponges (and therefore suspension feeders)(Hartman and Goreau, 1970) belonging to a little-known group that today participate in living coral reefs. The only possible non-suspension feeders in the Chazy reefs are the tabulate corals, presumably micro-carnivorous, and the trilobites, which are possibly detritus feeders (if not suspension feeders).

Maclurites is an archaeogastropod, and presumably grazed on algae. It is the largest of the primary consumers of the Chazy beds. The nautiloids may have fed on the Maclurites. If so, they ate the soft parts without breaking the shells, for most of the large shells are whole. The possibility that some of the nautiloids "grazed" on the sessile benthonic invertebrates should not be discounted, for these Middle Ordovician cephalopods are not far removed in evolution from the late Cambrian ellesmeroceratids, whose short, relatively non-buoyant shells indicate a vagile benthonic adaptation. Although the Chazyan nautiloids had buoyant shells and were probably good swimmers, they may have retained an interest in bottom feeding. Apart from the cephalopods we have no evidence for other large carnivores.

The Chazy reefs are thus a community in which the benthos were fed primarily from suspended matter or plankton. Even the corals had very small polyps and could not have eaten anything very much larger than a few millimeters across. This is surely a reflection of the paucity of larger nekton or vagile benthos on which to feed. Only the snails and cephalopods provide a larger fauna, and may have formed a side loop to the general food chain, the cephalopods feeding primarily on the snails.

Reefs earlier than the Chazy consist only of algae, or else include the demosponge Archaeoscyphia or possible sponges, such as

Archaeocyatha which were not likely to be anything other than suspension feeders. In the Silurian, corals become much more important, and one begins to see carnivorous macrophagy becoming a more important element in the trophic relationships of reef faunas. Silurian reefs are still dominated by tabulate rather than rugose corals among the carnivores, and thus consumed mainly small vagile animals. The suspension feeding element (bryozoa, stromatoporoids) is still strong in Silurian reefs. It is not until Devonian times that the large rugose corals become a dominant element in reefs, probably not without connection with the fact that this was the first time that fish and other large nekton appear in abundance.

DEVELOPMENT OF REEF FAUNAS

Within the Chazy Group the reef faunas show a progressive increase in diversity with time (Pitcher, 1964). The earliest reefs, in the lower Day Point Formation (Scott Member), are built of bryozoans only, and chiefly of one species, or at most two. These early benthic concentrations are, like their predecessors, of suspension feeders only. In the middle of the Day Point (Fleury Member) the *Lamottia* biostrome introduces the oldest-known coral in the world, which is also the oldest-known sessile carnivore with a skeleton. (The only older sessile carnivore is a possible anemone from the middle Cambrian Burgess Shale, and the only older vagile carnivores are the early Ordovician starfish, and the late Cambrian and early Ordovician nautiloids). It is possible at this moment in the history of the earth that it first became profitable for a carnivore to sit and wait for its food to come to it. This coral appears to have lived in a different environment from the bryozoa, although probably nearby. The corals are often fragmented and the fragments overgrown by bryozoa. Pitcher (1964, p. 648) considers the corals to have been transported into the area of outcrop, and there to have acquired their coatings of bryozoa.

Immediately above the *Lamottia* biostrome, in the upper Day Point, bryozoan mounds again develop (the "circular bryozoan reefs" of figure 1). On the surfaces of these circular mounds the laminar bryozoans that built them enclose large cystoid stems and small lithistid sponges, all in life-position and obviously part of a regular reef association (see plate 1, figure 1). These bryozoan-cystoid-sponge mounds foreshadow the richer fauna of the Crown Point reefs. Just beneath the Crown Point contact, there is another type of mound, built only of branching bryozoa (the "aligned bryozoan reefs" of figure 1). The colonies are relatively large and unbroken and seem to be preserved in place. An environmental difference (quieter water?) may account for this second type of mound.

In all the Day Point reefs algae are seemingly missing. In the Crown Point Formation the reef faunas are more diverse, and include algae, stromatoporoids, lithistid sponges, and corals

(*Billingsaria*, not *Lamottia*) along with the bryozoan species that built the earlier mounds. The algae were not immediately available to the reef animals for food, but probably performed an antipollutant function. They may have been eaten by soft-bodied meiobenthos which were subsequently consumed by the corals, but the principal flow of organic matter must have been from the phytoplankton directly, or through zooplankton, to the reef animals, and from there, through the intervention of bacteria, back to the benthic algae and to the phytoplankton as dissolved molecules, or recycled through the sponges in the form of whole bacteria, which may be a principal food of sponges (Rasmont, in Florkin and Scheer, 1968). That the Crown Point environment was in general one of high productivity is demonstrated by the abundance of the large snail *Maclurites magnus*, which is virtually a guide fossil to the formation, as well as of the large nautiloids that may have fed on it. The abundance of algae outside the reef environment (dead *Maclurites* and nautiloid shells are frequently encrusted by them) undoubtedly provided a firm base for the overall food chain as well as a food supply for the *Maclurites*.

In the Valcour Formation the faunal complexity is maintained and the reef assemblage differs little from that of the Crown Point.

In the Champlain Valley and elsewhere the Crown Point reefs represent the earliest appearance of a complex reef-building community. It is worth noting that the animals involved (stromatoporoids, lithistid sponges, tabulate corals, and bryozoa) are very nearly the earliest representatives of their respective taxonomic groups. In large part this new complexity is due to the evolution of new life. Earlier reefs, even the most complex known, such as the early Ordovician mounds of Texas (Toomey, 1970), are built of only a few organism types, usually algae and sponges.

The abundance of sponges in the Crown Point and early Valcour reefs deserves consideration, for it can be related to the general evolution of hermatypic organisms. Sponges are not common reef-building animals. During the period when tabulate and rugose corals were abundant, and during the periods when the scleractinians were abundant, including today, sponges were a very minor element in the construction of reefs. It is only before the corals first become abundant (before the Silurian), and also during the interval between the decline of the Paleozoic corals and the rise of the Scleractinians (Permian and Triassic), that sponges were important reef builders. This statement leaves out the stromatoporoid sponges, which managed to coexist with the Paleozoic corals through the Silurian and Devonian, though often in different reefs, and presumably in different environments. The bryozoans show a similar relationship to the corals but seem to have been sturdier competitors than the sponges. Bryozoa are still present in Silurian reefs alongside corals and stromatoporoids, though they tend to be replaced by the latter in ecologic successions (Lowenstam, 1957). In Devonian times bryozoans are

rarely present in reefs, but reappear in the Carboniferous and Permian (Zechstein of Germany) when corals declined.

The Ordovician reef-sponges are of interest in that they are siliceous (lithistid demosponges) rather than calcareous (although frequently calcified diagenetically). They first appear (Archaeoscyphia) in dominantly-algal early Ordovician mounds (Toomey, 1970) but become more abundant and diversified in Middle Ordovician (Chazy and Black River) reefs, which is the only time during the Paleozoic that siliceous sponges were significant frame builders. This time coincides with the first radiation of the lithistid demosponges. It may be that the higher rate at which stromatoporoids, corals and bryozoa could secrete calcium carbonate skeletons was the reason for the near disappearance of lithistids from the later reefs. When the corals declined at the end of the Paleozoic, it was the calcareous Sphinctozoan sponges that replaced them as important reef-builders, in Permian and Triassic times, along with the ever present calcareous algae.

It should be noted that most lithistid sponges, although their skeletons are rigid, do not by themselves bind sediment or build up massive structures. In the Crown Point reefs sediment binding was probably carried on only by stromatoporoids, laminar bryozoa, corals, and calcareous algae. Nevertheless, the lithistids cover, on the average, from 22% to 50% of the surface of the reefs in which they are most abundant (Pitcher, 1964, p. 662, 675). They thus contributed significantly to the bulk of the reef mass. They also served to trap sediment. That this by itself can be a potent factor in mound formation is indicated by the late Jurassic sponge "reefs" of Germany (Roll, 1934) in which siliceous sponges built mounds apparently solely by trapping sediment and without the significant presence of binding organisms. These Jurassic mounds are the only known examples of siliceous sponge reef-like structures in post-Paleozoic times.

It is possible that some of the laminar Anthaspidella was actually of encrusting habit and may have helped to bind other skeletal material, but its role would have been minor compared to that of the more abundant binding organisms.

ECOLOGIC SUCCESSION

Ecologic succession in the Day Point reefs can hardly be said to exist, since the reefs consist only of one species of bryozoan. The encrusting of the coral Lamottia by the bryozoan Batostoma is probably not true succession if the corals are not in place. It is worth noting, however, that the Lamottia bed is immediately succeeded by Batostoma reefs which were built on the coral debris (Pitcher, 1964, p. 650) as shown by cores.

In the more complex Crown Point reefs no clear succession is evident though there are suggestions of it. The stromatoporoid *Pseudostylodictyon* frequently forms small reeflets by itself, resting on pelmatozoan calcarenite. It also often forms the basal parts of larger reefs, together with subordinate ramose bryozoa. Subsequently there succeeds a more diverse fauna of the lithistid sponges *Zittelella* and *Anthaspidella*, the coral *Billingsaria*, the bryozoan *Batostoma* and a flora of *Sphaerocodinium* and *Solenopora*. Some reefs on Valcour Island end with this community. Others on Isle La Motte often have a capping of *Pseudostylodictyon* alone. In this mature reef community the lithistid sponges and the stromatoporoids occupy by far the largest surface area. *Billingsaria*, bryozoans and the algae are distinctly subordinate. The stromatoporoids can be considered to form a pioneer community which initiates reef development. It apparently provides a favorable substrate for the lithistid sponges and for the encrusting corals, bryozoans and algae. The lithistid sponges (*Zittelella*, *Anthaspidella*) can be quite common in the calcarenites away from the reefs, and therefore do not need the stromatoporoids as a base. Their participation in the reef is facultative rather than obligatory. The development of this rudimentary succession may be a matter of building up into somewhat shallower water, as is suggested by the change from ramose to laminar algae. It may also be a matter of the development of a firmer substrate than is provided by the surrounding shell sand. Biotic factors such as the availability of food probably also enter into the picture. Laminar algae, favored in their growth by a hard substrate, may attract herbiverous meiobenthos, which may in turn provide abundant food for *Billingsaria*, and indirectly, more bacteria for the sponges.

In the Chazyan mounds of Quebec, a better-defined ecologic succession has been ascertained (see Toomey and Finks, 1969). Here pioneer communities of the encrusting bryozoan *Batostoma* are succeeded by a mixed bryozoan-coral community (*Batostoma*, *Chazydictya*, *Billingsaria*, *Eofletcheria*). Finally the corals (*Billingsaria* or *Eofletcheria*) become dominant over the bryozoans at the top of the mound, perhaps a foretaste of things to come.

COMPETITION

Bryozoans tend to show a somewhat inverse relationship of abundance with reference to stromatoporoids and sponges (Pitcher, 1964, figure 44) suggesting competition, as might be expected from the fact that they are all suspension feeders. At the top of the Crown Point, bryozoan reefs occur side by side with stromatoporoid-lithistid reefs. They tend to dominate the Valcour reefs again, almost as they did in the earlier Day Point. The variability of the proportions of reef organisms in the Crown Point from one reef to the next, also suggests that there was near-equality

in competition between many of these organisms. At least one reef in the pasture on Isle La Motte is composed of 50% Billingsaria throughout (Pitcher, 1964, p. 666). Other reefs in the same pasture contain, on the surface, anyway, about 50% lithistid sponges (Zittelella, Anthaspidella). The corals and the sponges did not compete for food but they probably competed for substrate space. Occurrences of Billingsaria and Zittelella together on the flanks of reefs in this same pasture indicate that they had the same environmental tolerances.

VERTICAL ZONATION

In the early Day Point bryozoan mounds, the mounds are built of laminar Batostoma or Cheiloporella, while the interreef areas contain abundant branching Atactotoechus. This may be considered a rudimentary sort of depth zonation, with the branching bryozoa occupying the deeper quieter water, and the laminar bryozoa the rougher shallower zones. However, the total relief at any one time was scarcely more than a foot or two (see Pitcher, 1964, figure 10) and the differences in wave energy could not have been very great. Nevertheless, the presence of branching bryozoa, along with stromatoporoids, in the basal parts of Crown Point reefs, and their replacement by laminar bryozoa higher up (Pitcher, 1964, figure 8) suggests that there may be something to this form distribution in relation to depth. Certainly in living sponges, corals, and bryozoa there is a similar confinement of branching forms to the less rough water areas.

The surface distribution of organisms on a Crown Point mound was studied by Pitcher (1964, figure 26) from the low flanks up to its crest. This should reflect bathymetric differences. He found that the stromatoporoids were most abundant at the crest, the bryozoa most abundant somewhat lower down, and the corals and lithistid sponges most abundant still lower on the flanks with the sponges remaining abundant further down than the corals. This again would correspond to a well-known pattern of morphological distribution, with the conical or cup-shaped lithistids (Zittelella) being characteristic of quieter, deeper water, while the laminar bryozoans, corals and stromatoporoids are characteristic of rougher water. The total vertical relief involved is scarcely six feet, and except for the absence of stromatoporoids at the base and the absence of lithistids and bryozoa on the crest, all the forms occur over the whole reef. Thus the environmental differences cannot have been very great.

A more pronounced bathymetric differentiation may be shown by some of the Crown Point reefs on the southwest shore of Valcour Island, on the point of land north of the concrete boat dock. Here the flanking beds pass laterally into dark calcilutites with numerous hexactinellid sponge root-tufts and body fragments. These

are much more delicate sponges and may have occupied a depressed area with genuinely quiet water peripheral to the reef.

ORIENTATION AND CURRENTS

Bryozoan mounds in the Day Point (Pitcher, 1964, figure 19) and stromatoporoid reeflets in the Crown Point (on both Isle La Motte in the Goodsell Quarry, and on the mainland at Sheldon Lane) tend to have a roughly north-south orientation. This is parallel to the paleoshore, and the mounds may have grown either in belts of optimum depth or into the set of longshore currents. An indication that currents may be involved is shown by the fact that hexactinellid sponge root-tufts in non-reefy beds of the Crown Point at South Hero, Vermont (Pitcher, 1964, figure 32) show the same preferred orientation on the bedding planes. Orthocone nautiloid shells are less clearly oriented, but in the channels that cut the Crown Point reefs, nautiloid shells are most commonly oriented parallel to the axis of the channel, obviously parallel to currents sweeping through. Maclurites shells are also often piled together in pockets in these channels, probably as a result of current action. The channels, however, may not be strictly contemporary with the reefs they cut.

CHANNELS

The Crown Point reefs are cut by numerous channels, mostly one to three feet wide and as much as two feet deep, filled with a black calcarenite that contrasts sharply with the light calcilutite of the reef rock. There is a considerable body of evidence that these channels may have been formed subaerially by solution, originally pointed out by Oxley and Kay (1959, p. 831), possibly by enlargement of tectonic joints, following consolidation and diagenesis of the reef rock. The entire sequence of events would have to have taken place entirely within Crown Point time, perhaps several times. The evidence is as follows:

1. The channels have sharp boundaries against the reef rock along smooth surfaces that cut through the middle of stromatoporoid colonies, lithistid sponges, and calcilutite matrix in a continuous sweep. The matrix must have been consolidated, and the lithistid sponges may have already been changed from silica to calcite, for they show no effects of differential hardness on the erosion surface.

2. The channels usually end in rounded culs-de-sac, or sometimes have an ovoid shape, suggesting either pothole-like abrasion or sinkhole-like solution. There are essentially no quartz clasts in the surrounding sediments, so that abrasion would seem to be unlikely, thus leaving solution as the alternative.

3. The channels tend to intersect at close to right angles and most frequently, though by no means universally, are oriented roughly north-south and east-west. This suggests that they may follow a tectonic joint pattern. Participants in the trip are invited to compare the form of the channels with that of solution-enlarged joints now being eroded in the same rock.

4. If the channels were surge channels present in the active reef, we would expect to find them bordered with at least some entire outlines of reef-building organisms, or where these were broken by contemporary wave-action, to find the broken outlines, and margins of the channel as a whole, to be irregular rather than smooth.

Because the calcarenite filling the channels contains Crown Point guide fossils identical to those beneath and to either side of the reef, and because such channeled reefs occur at more than one level within the Crown Point beds in the same area, we must assume that the entire process postulated took place repeatedly within Crown Point time. If Crown Point time is assumed to be one-third of Chazy time, and that one-sixth of Ordovician time and Ordovician time to be 60 million years long, then we have 3.3 million years for these processes to take place in. Admittedly, this may be hard to swallow, and we have not had the opportunity to test the hypothesis adequately, but participants in the field trip may wish to think about these possibilities while examining the outcrops. Gavish and Friedman (1969) have recently demonstrated post-glacial (within 10,000 years) calcification of quartz sand grains during consolidation of later glacial eolianites under subaerial conditions, thus providing strong support to this hypothesis. On the other hand there are elongate calcilutite-filled pockets, often containing numerous large nautiloid shells, that occur within the reef mounds of both Crown Point and Valcour. These may have been channels contemporaneous with the reefs. In some, algal coatings cover the shells and also line the walls of the pockets (see Goodsell Quarry, for example).

ITINERARY

The walking tour will start at the north end of the picnic ground and trailer camp on the north side of Wait Bay in southeastern Isle La Motte. It may be reached by following the main north-south road down the center of Isle La Motte to its southern end, turning left (east) to the trailer park entrance, and then turning left (north) up the hill to the picnic ground. Please note that the entire trip is on private property, and that permission must be secured from the landowners for visits.

Cross the fence and walk north to the bare exposures of the

Lamottia biostrome in the Fleury Member of the Day Point Formation. CAUTION ! DO NOT STEP INTO SOLUTION-ENLARGED JOINTS. SOME ARE PARTLY CONCEALED BY VEGETATION. WALK ONLY ON BARE ROCK SURFACES. THE JOINTS ARE OVER A FOOT DEEP.

The hemispherical to discoidal heads of Lamottia are closely packed in a calcarenite matrix. Joints offer an opportunity to observe their orientation in section. More than half are overturned over much of the area. Many are broken. The proportion of broken ones increases to the north and east, where the biostrome passes into calcarenite with ever fewer and smaller fragments of Lamottia. In the central area of the exposure there are belts some 10 feet wide in which fragmentation, proportion of overturned specimens, and quantity of calcarenite matrix, are higher than elsewhere. These may represent surge channels. In the peripheral area to the northeast one may see much laminar Batostoma chazyensis surrounding the Lamottia fragments.

This is the type locality for the genus Lamottia Raymond (1924). Although Raymond's description of this bed as the "world's oldest coral reef" may be disputed, it still seems to be unchallenged as the world's oldest occurrence of corals of any kind.

Walk northwestward upsection, so far as fence lines, cultivated fields and vegetation permit. DO NOT DISTURB FENCES OR LEAVE GATES OPEN ! NO SMOKING WHILE WALKING THROUGH THE FIELDS; THERE IS A DANGER OF FIRE. ALSO, PLEASE KEEP OFF CULTIVATED GROUND.

About 1400 feet to the west are circular mounds of laminar Batostoma chazyensis containing small sponges near their periphery as well as cystoids in place throughout. These mounds (see Plate 1) immediately overlie the Lamottia bed. Another 1000 feet to the north brings us to elongate mounds (aligned N-S) of branching bryozoa at the very top of the Day Point Formation. This may represent deeper water.

Continue to walk northward to a small dirt road, then walk west along it to a T-junction with a larger dirt road. Turn left and follow it southwest to a house and barn on the right. We will enter a gate into the large pasture behind the house and barn. Mr. Ira LaBombard, the present owner of the property, has kindly given us permission to enter his pasture to study the reefs in the Crown Point and lower Valcour Formations. He has requested, as a condition of this permission, that NO SPECIMENS WHATEVER be collected. PLEASE RESPECT THIS ORDER !!! We will have an opportunity later in the day to collect from these same beds at another locality. The fossils are so beautifully displayed here that relationships may be seen without disturbing the rock. They may be photographed very advantageously on the glacially polished surfaces.

The reefs exposed here are mainly in the Crown Point Formation and are the ones intensively studied by Pitcher (1964).

You may examine contemporaneous reefs by walking northeastward along strike. You may examine younger reefs by walking northwestward upsection (dip is about 10 degrees NW).

The reefs are exposed as mounds of light rock. The calcarenite between the reefs, and filling the channels in the reefs, is nearly black. The reefs outcropping nearest the fence were mapped by Pitcher as his assemblage A, consisting of the stromatoporoid Cystostroma and the alga Solenopora. Those beyond to the northwest, and covering most of the pasture up to a distinct linear rise in the ground, belong to Pitcher's assemblage B. These show interesting variations from reef to reef as well as changes in faunal distribution from flanks to tops of the mounds. The fauna consists of the stromatoporoid Pseudostylodictyon eatoni, the lithistid demosponges Zittelella varians and Anthaspidella sp., the tabulate coral Billingsaria parva, the bryozoan Batostoma chazyensis, and the calcareous algae Solenopora, Sphaerocodinium and Girvanella.

The fossils may be identified readily on weathered surfaces as follows:

1. Pseudostylodictyon eatoni: Large whitish masses with fine dark laminae forming concentric patterns about centers an inch or two apart. These concentric patterns represent the mamelons and their small size is characteristic of the species.

2. Zittelella varians: Circular, dark gray bodies two to three inches in diameter, with a central circular light area representing the matrix-filled cloaca, and radial light areas, or ovoid dots, a few millimeters wide, representing the canals. In longitudinal section, the sponge is conical, and oblique sections will show the expected intermediate shapes. Some specimens have an irregular outline in cross section.

3. Anthaspidella sp: Similar to Zittelella in color and texture, but shaped like long sinuous bodies, and inch or so thick and several inches long, when seen in cross section. A surface view of the sheet-like sponge shows a somewhat irregular mass without a cloaca. The complete sponge has a short stalk, the whole being shaped somewhat like a distorted cake-plate. The open 'spongy' texture may help when shape fails. Needless to say, the shape and geometric arrangement of the spicules in thin section is necessary for a secure identification. Not every shapeless mass is a sponge.

4. Billingsaria parva: Small, black, oval patches, a few inches across. The dark color is very distinctive. Close inspection with a hand lens will reveal the stellate outlines of the corallites with their characteristic septal ridges.

5. Bryozoa: These weather white, either as small branching twiglets, or as laminated sheets. Identification requires thin-sections, but the outlines of the zooecia are usually visible on the weathered surface and suffice to identify it as a bryozoan.

6. Solenopora: White concentric circles, often sparry. A few inches across. This is the most common form of Solenopora seen on the reef surfaces.

7. Girvanella: Small black ovoid bodies, less than an inch in length. These are oncolites, or algal-coated shell fragments.

8. Maclurites magnus: Large coiled shells a few to several inches across. No septa. The shell substance is white in cross-section.

At the rise in ground is a one-foot stromatolitic layer with many orthocone cephalopods. Pitcher called this his assemblage C and assumed it was laid down as a blanket during a relative drop in sea level. It forms a dip slope through which appear, apparently, the tops of assemblage B mounds, as well as small mounds of Batostoma chazyensis alone which Pitcher called assemblage D. At the west end of this cuesta-like feature, nearest the main road, a good cross-section of an assemblage B mound is exposed (see Plate 2).

Down the dip slope, above a ten-foot interval of grey calcarenites, are mounds in the lower part of the Valcour Formation. They are composed of Batostoma campensis, together with the alga Solenopora. Some Zittelella may be found. The bryozoa are clearly dominant.

Walk northeastward along strike for about a half-mile, observing Crown Point mounds as you go. You will eventually reach the Goodsell Quarry, operated by the Vermont Marble Company. The quarry is opened in the lower beds of the Crown Point which are relatively lacking in reefs except for small stromatoporoid-algae mounds. The quarry has been intermittently active, and the stone, which makes a beautiful black marble when polished, has been widely used as an interior trim. The rock weathers light gray, and has also been locally used as a dimension stone. It was used to build the old fort, Fort Montgomery, visible from the Rouses Point Bridge.

CAREFULLY avoiding falling into the water-filled quarry, one may observe vertical sections through stromatoporoid-algal mounds and their relationships with the surrounding calcarenite (see Plate 5). By tracing laminae from the mounds into the surrounding sediment, one can see that the mounds never stood more than a foot or two above the sea floor at any one time, though the total thickness is much greater because of the persistence of the mound population

on the same spot. On the quarry benches, especially the glacially polished upper surface, one may see plain views of mounds and note their tendency to a N-S lineation. On these surfaces also, especially when wet down, one may see orthocone and other nautiloid shells, and Maclurites shells, overgrown by algal coatings.

ACKNOWLEDGEMENTS

We wish to thank Mrs. Malvina Bruley and Mr. Ira LaBombard of Isle La Motte, for permission to visit the classic exposures on their respective properties. Their cooperation has made this excursion possible. R. M. Finks extends special thanks to Mr. Rodney V. Balasz for information concerning possible channels in the Lamottia biostrome, and to his paleontology class for mapping some of the channels in the various reefs in the fall of 1968.

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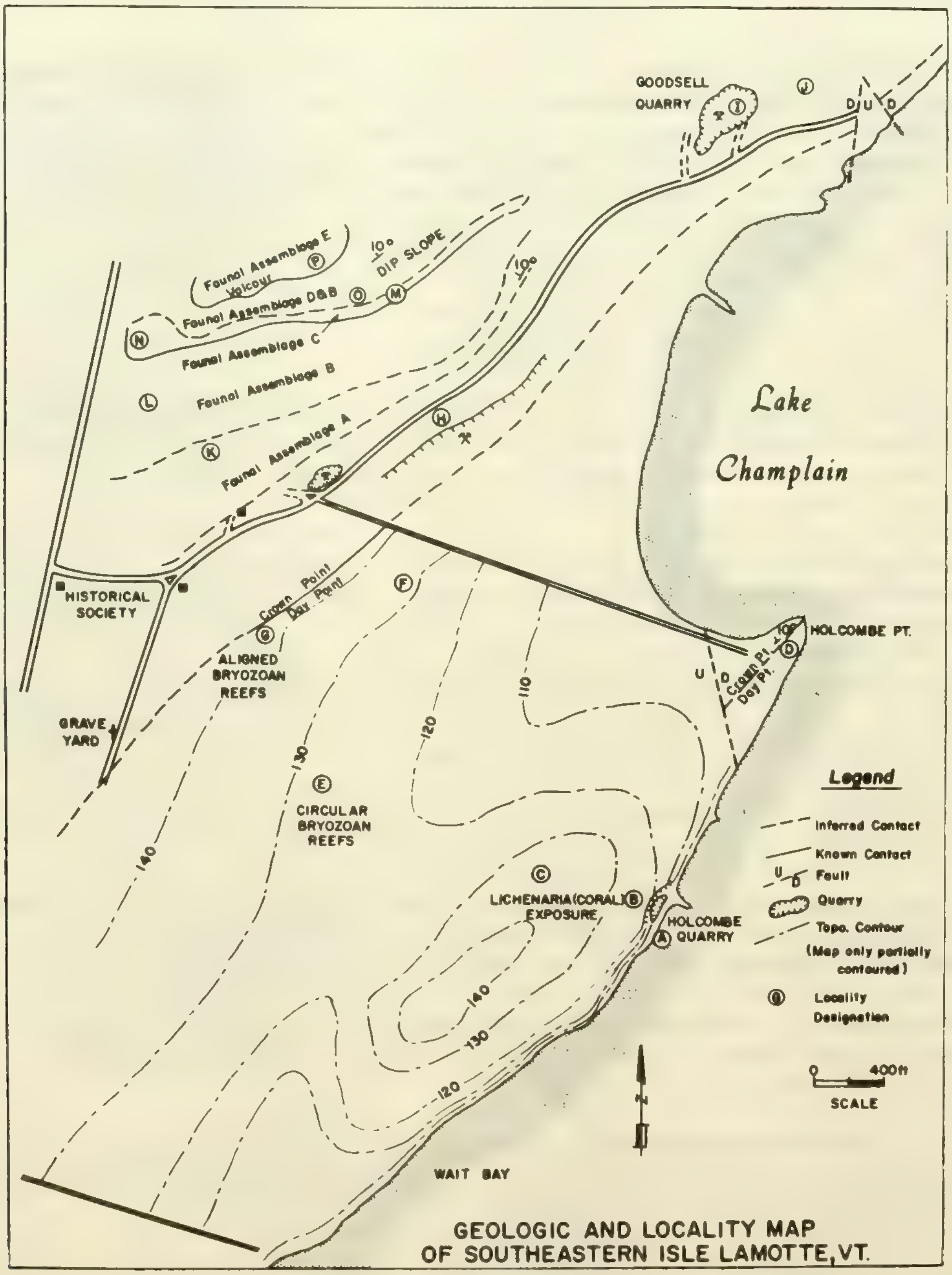

Figure 1. Geologic and locality map of Southeastern Isle La Motte, Vermont. After Pitcher, 1964.

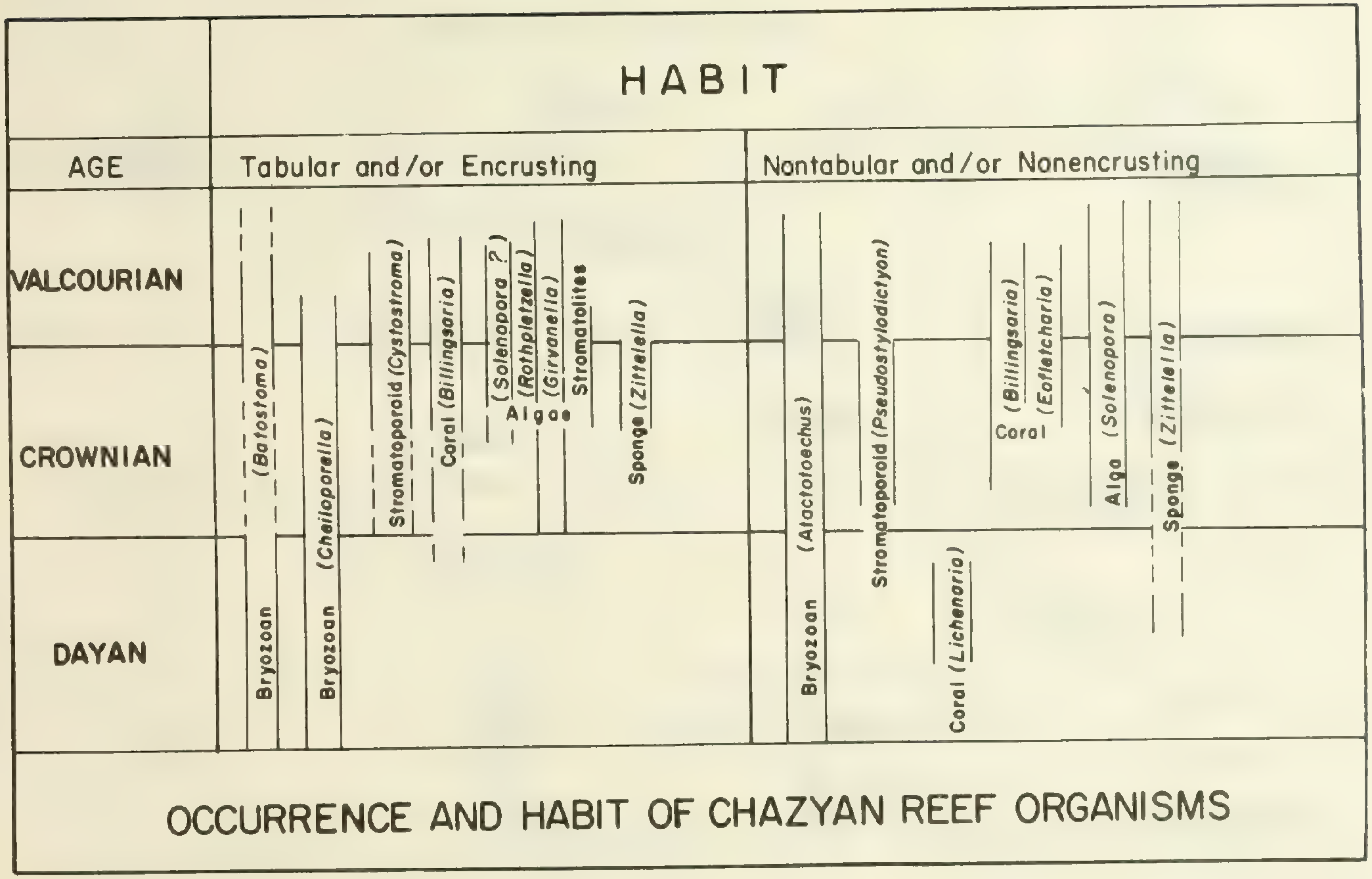

Figure 2. Occurrence and habit of Chazyan Reef organisms. After Pitcher, 1964.

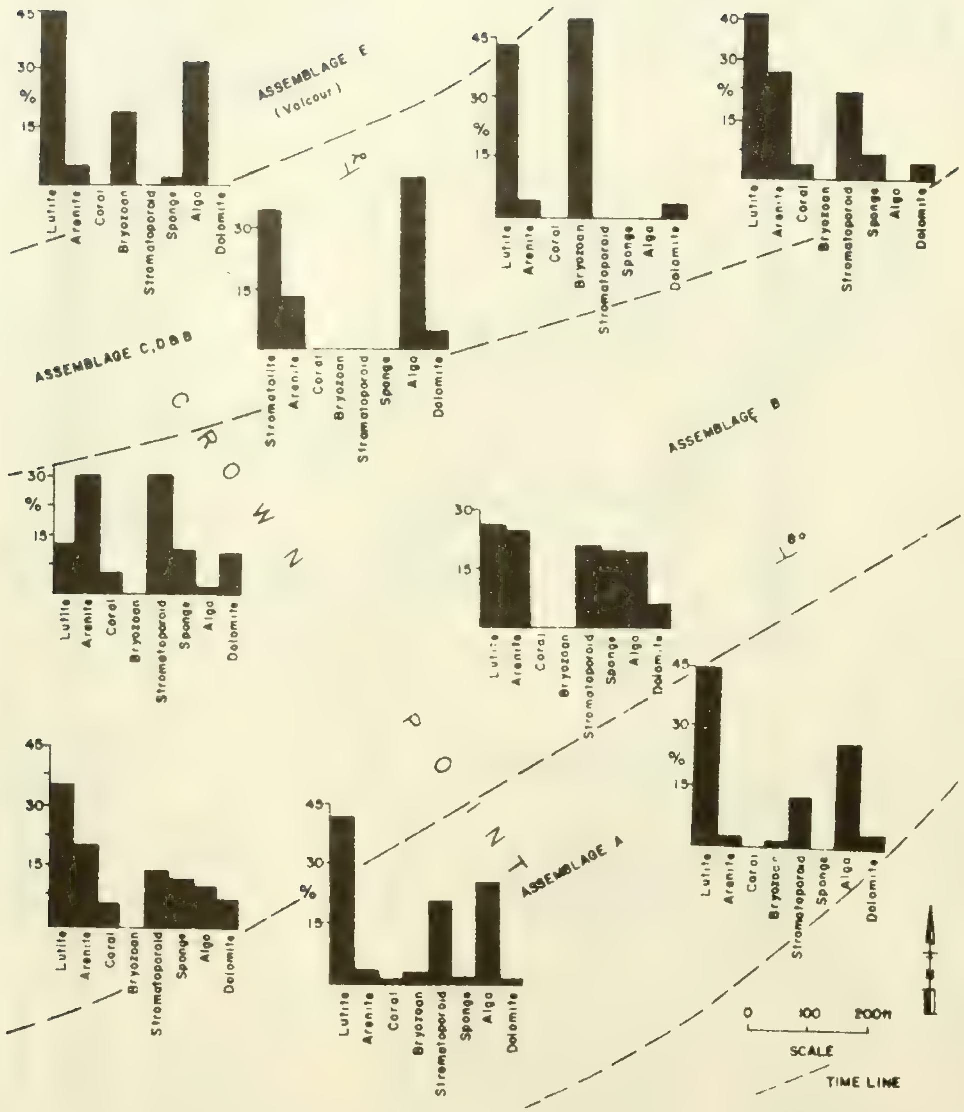

Figure 3. Map showing composition of several reef assemblages in LaBombard's pasture, Isle La Motte. Vermont.(from Pitcher,1964)

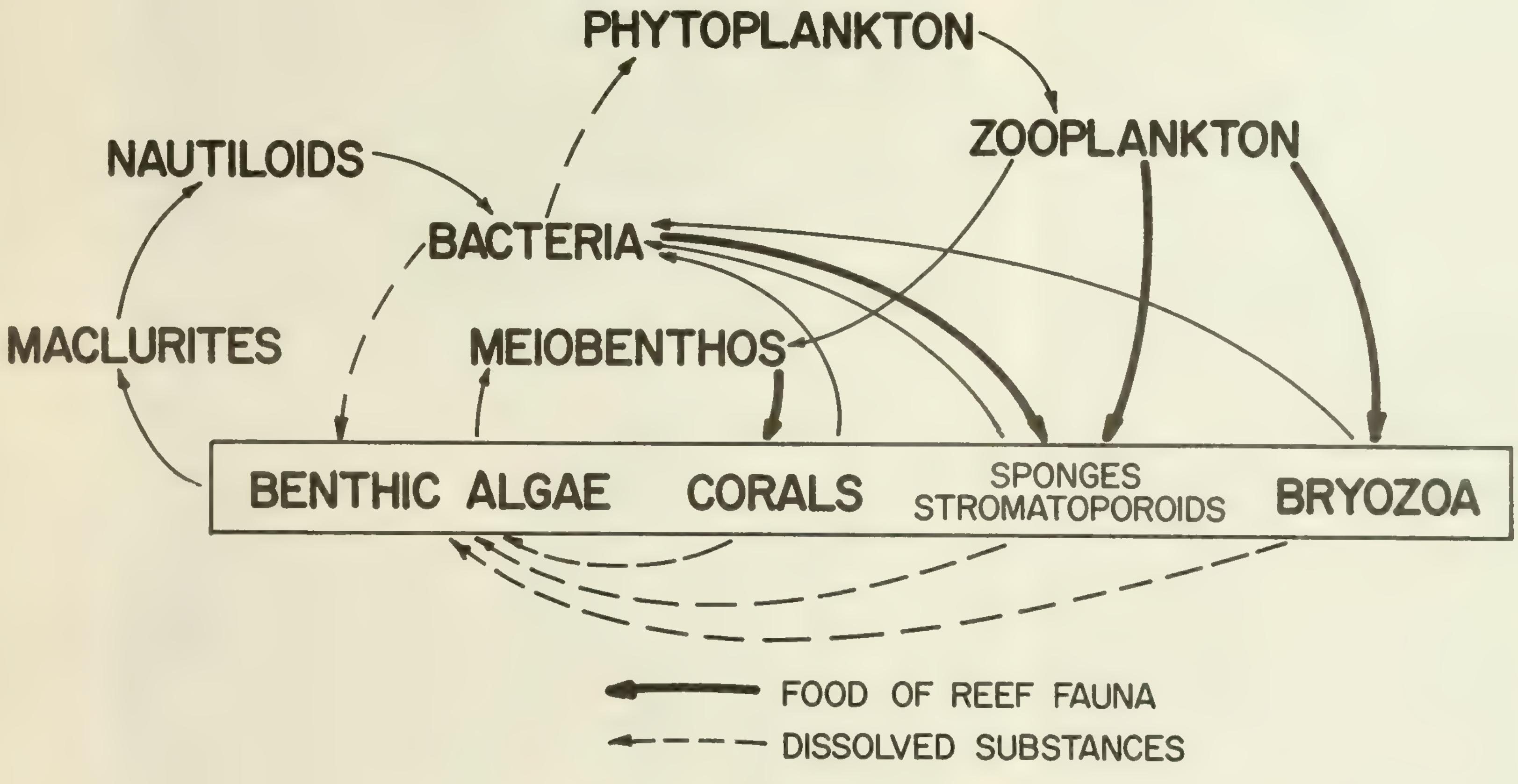

FIGURE 4. SIMPLIFIED FOOD-CHAIN OF CROWN POINT REEFS

Figure 4. Simplified food-chain of Crown Point reefs (after Finks and Toomey, 1969).

PLATE 1

Figure 1 Outcrop photograph of the surface of a Middle Ordovician (Chazyan) bryozoan mound exposed in the Day Point Formation (uppermost Fleury Member) on Isle La Motte, western Vermont. Note general lineation of the bryozoan colonies and included round sponges; length of hammer approximately 14 inches.

Figure 2 Thin section photomicrograph (X3) of characteristic bryozoan mound rock that forms conspicuous mounds in the uppermost Day Point Formation, Fleury Member, on Isle La Motte, western Vermont. The mound rock is primarily composed of consecutive sheets or layers of the colonial trepostome Batostoma chazyensis Ross, separated by lime mud layers containing relatively abundant, although quite small, dolomite rhombs (small grey flecks).


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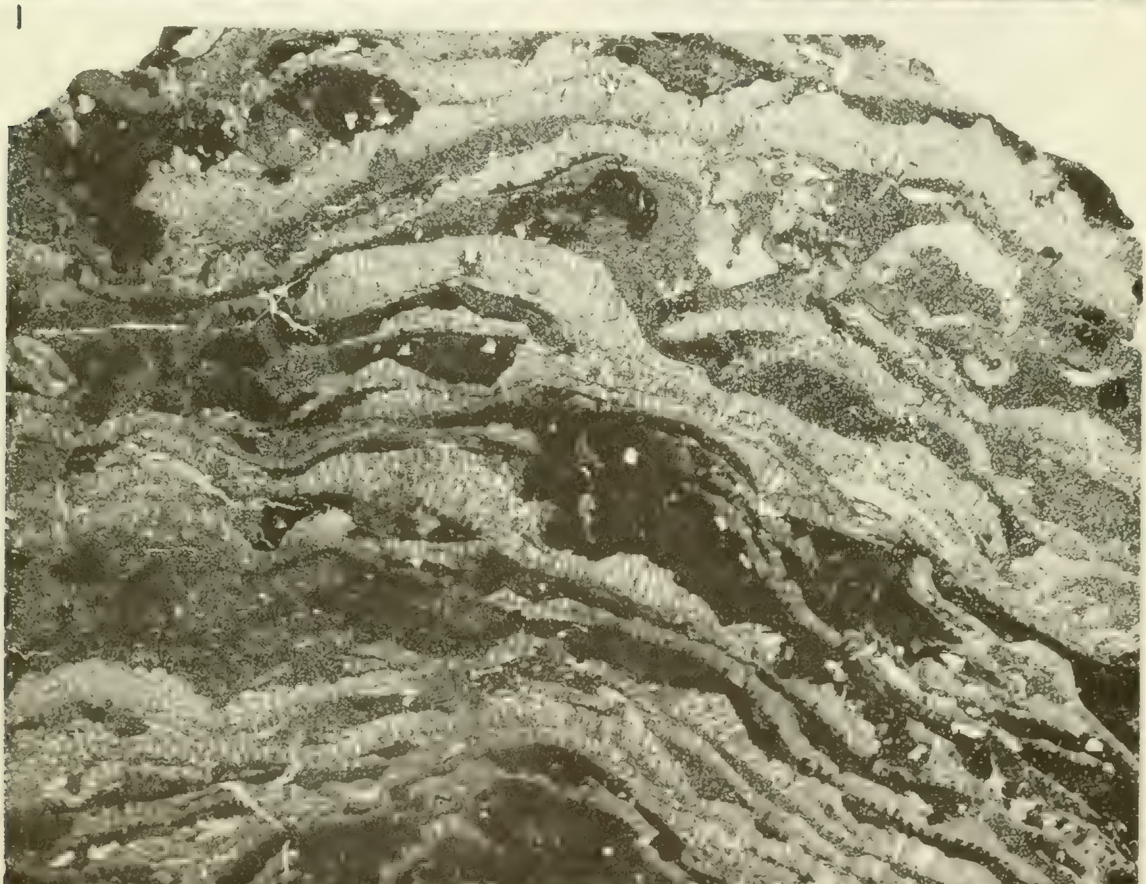
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PLATE 2

Outcrop photograph of a series of typically small-sized Crownpointian (Middle Ordovician - Chazyan) mounds exposed in LaBombard's pasture, Isle La Motte, western Vermont. Rounded mound structures are composed of lime mud containing abundant algae, sponges, stromatoporoids and trepostome bryozoans. The beds filling in the surface irregularities and capping the mounds are dominantly composed of relatively coarse-textured pelmatozoan debris (see Plates 3 and 4). The length of the sledge hammer located on the right hand side of the prominent mound is approximately 3 feet.

PLATE 3

Figure 1 Thin section photomicrograph (X3) of a transverse cut through the sponge _Zittelella_ in what is typically Crownpointian mound rock; LaBombard's pasture, Isle La Motte, western Vermont. Note overall muddy character of the rock, and the appearance of an encrusting bryozoan ? on the outer surface of the sponge.

Figure 2 Thin section photomicrograph (X14) of Crownpointian mound rock with relatively abundant encrusting (bead-like segments) algae of the genus _Sphaerocodium_; LaBombard's pasture, Isle La Motte, western Vermont. Again, note the dominantly muddy character of the rock. Scattered small grey flecks are floating dolomite rhombs.

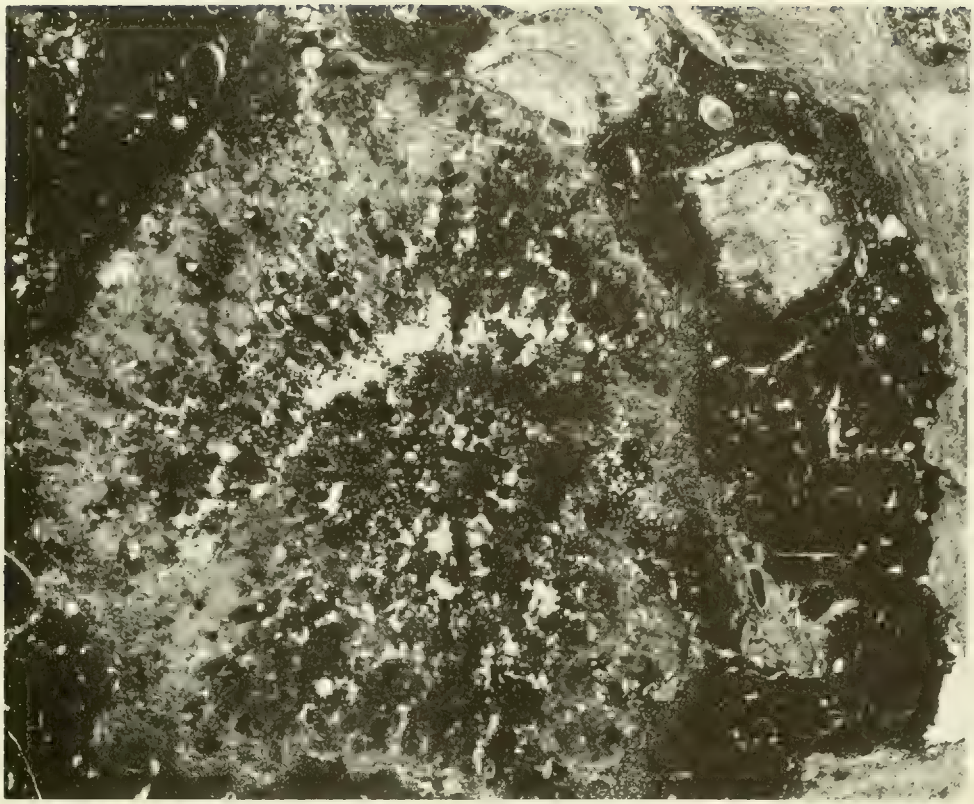

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PLATE 4

Figure 1 Outcrop photograph of a channel cutting mound rock in the Middle Ordovician (Chazyan) Crown Point Formation, west of the Goodsell Quarry, Isle La Motte, western Vermont. The width of the channel is approximately 18 inches. Note lighter colored mound rock on either side of darker-colored channel rock.

Figure 2 Thin section photomicrograph (X4) of channel rock from the above locality. Rock is primarily a pelmatozoan calcarenite, although intraclasts (small rounded dark grains), bryozoan and brachiopod fragments are also present. Cavities or original void spaces filled with secondary granular sparry calcite are also common within the channel rock.


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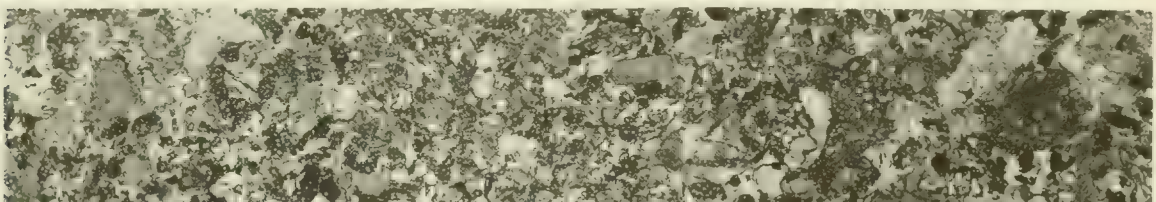
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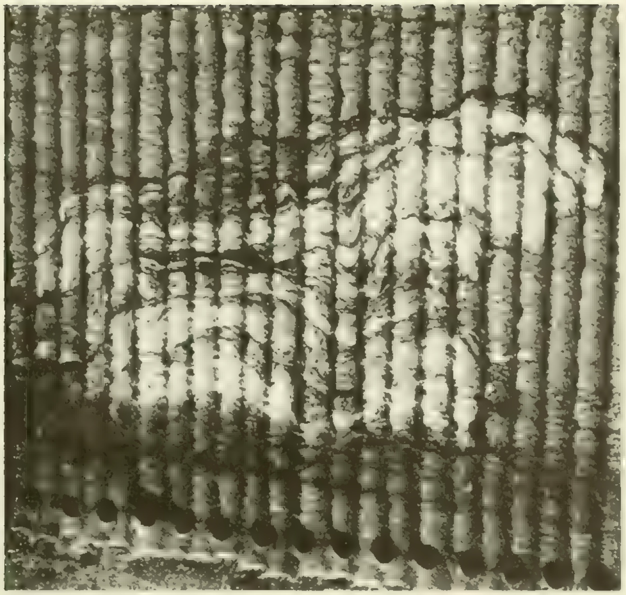
PLATE 5

Figure 1 — Stromatoporoid (Pseudostylodictyon ?) mound exposed on the south wall of Goodsell Quarry (June, 1962); Middle Ordovician (Chazyan) Crown Point Formation, Isle La Motte, western Vermont. Stromatoporoid mound is approximately 4 feet in width and 2 1/2 feet in height.

Figure 2 — Thin section photomicrograph (X4) of a vertical section of the stromatoporoid Pseudostylodictyon ? chazianum (Seely) from the lower part of the Crown Point Formation, LaBombard's pasture, Isle La Motte, western Vermont. Specimen shows characteristic thin laminae separated by pronounced layers of lime mud.

Figure 3 — Thin section photograph (X4) of a horizontal section of the stromatoporoid Pseudostylodictyon ? eatoni (Seely) showing mamelons of various sizes, from the lower part of the Crown Point Formation, LaBombard's pasture, Isle La Motte, western Vermont.



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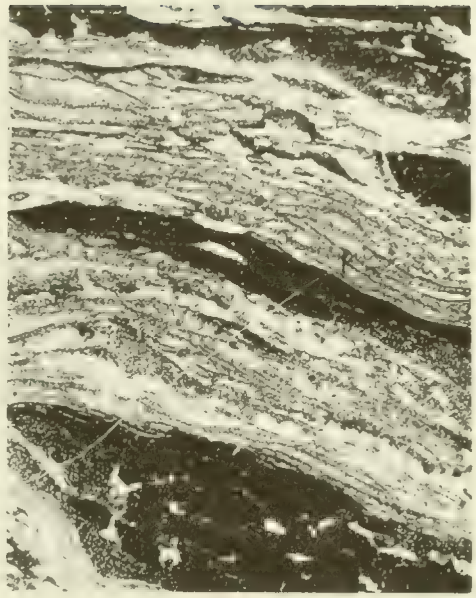

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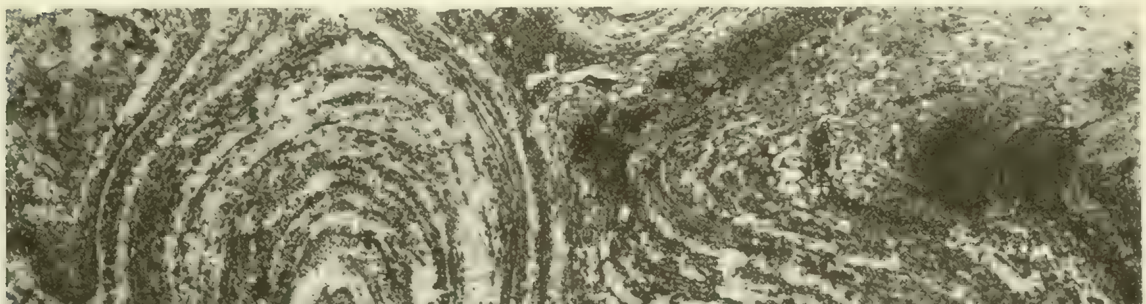

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PLATE 6

Figure 1 Thin-section photomicrograph (X8) of a pelmatozoan grainstone from a channel associated with the Lamottia buildup ("Lamottia reef" of Raymond), Day Point Formation (Chazyan) middle Fleury Member, Isle La Motte, western Vermont. Note abundance of pelmatozoan ossicles (probably cystoid and/or blastoid), and the dominant sparry calcite matrix; many of the pelmatozoan ossicles have calcite overgrowths. The small black grains are intraclasts and/or Girvanella pellets (diagnostic structures not seen at this magnification).

Figure 2 Outcrop photograph of the Lamottia accumulation ("Lamottia reef") located near the center of the buildup. Note the jumbled mass of coral "heads" which appear to be heaped together and overturned, probably due to wave and/or current sorting. Length of hammer is 11 inches; Day Point Formation, middle Fleury Member, southeastern Isle La Motte, western Vermont.

Figure 3 Thin-section photomicrograph (X4) of the muddy rock matrix between the massive Lamottia "heads". Rock can be classified as a skeletal wackestone. Note large fragment of the tabulate coral Lamottia heroensis Raymond set within a muddy matrix with included intraclasts and abundant skeletal debris. Thin-section taken from rock near the center of Raymond's "oldest coral reef" within the Day Point Formation, middle Fleury Member, southeastern Isle La Motte, western Vermont.

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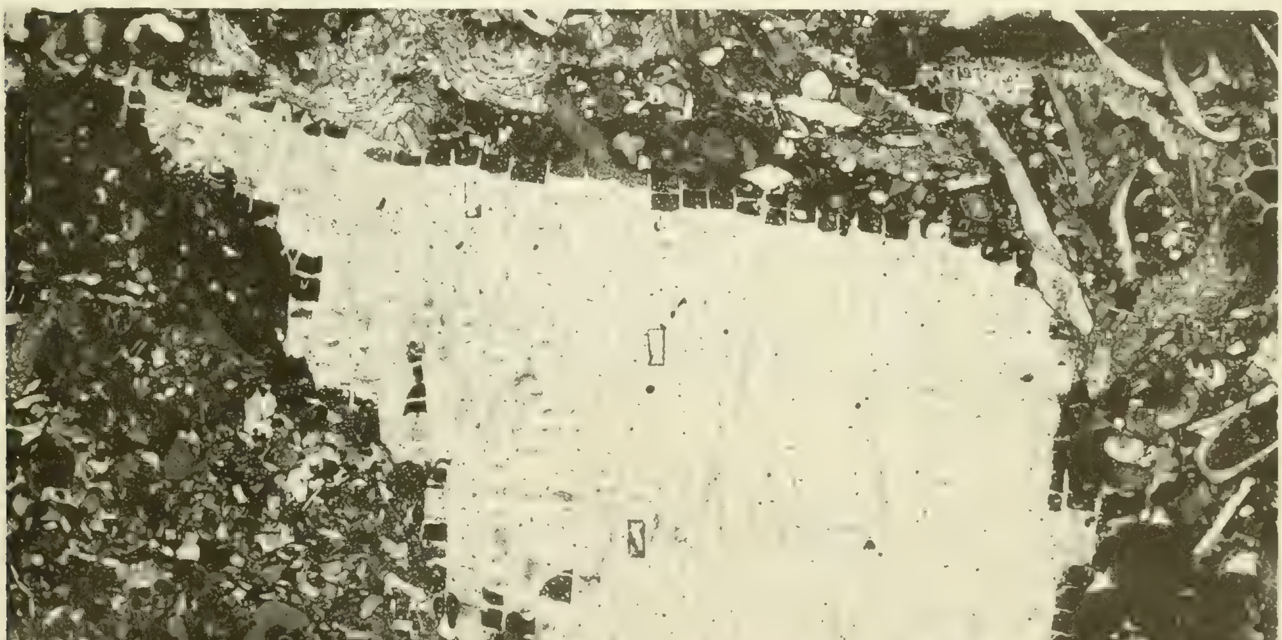

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Trip P-2

CAMBRIAN FOSSIL LOCALITIES IN NORTHWESTERN VERMONT

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Rutgers University
Newark, New Jersey

INTRODUCTION

The Cambrian sections of northwestern Vermont are widely regarded as classic on account of the discovery therein of Cambrian fossils by Zadock Thompson and S.R. Hall in 1847 and Noah Parker in 1954, as well as on account of the variety of the fossils subsequently collected. The first fossils found were described by Elkanah Billings and James Hall. Further discoveries were made by Charles Walcott. More recent paleontological work includes that of Clark and Shaw (1968a; 1968b), Howell (1937), Kindle and Tasch (1948), Rasetti (1946), Raymond (1924; 1937), Resser and Howell (1938), Schuchert (1937), Shaw (1951-1966), and Tasch (1949).

Recent regional syntheses (Cady, 1968; Palmer, 1971; Rodgers, 1968; Theokritoff, 1968) have interpreted the Cambrian rocks of northwestern Vermont as representing, on the one hand, the deposits of a sand-carbonate shelf extending to the west and northwest onto the craton and, on the other hand, the deposits of a deeper water basin situated to the east and southeast. The sand-carbonate shelf formed a steep bank, rising above the basin and contributing carbonate clasts to sites of dominantly shale deposition at the foot of this sand-carbonate bank. The boundary between the shelf facies and the basin facies is now complicated by and partly obscured by thrusting.

The present trip will visit three fossil localities. The first locality is in the Lower Cambrian Monkton Quartzite, the second in the Lower Cambrian lower Parker Shale, and the third in the Upper Cambrian Gorge Formation. The first and the third are in the shelf facies, the second in the basin facies.

It is a distinct pleasure to acknowledge the courtesy of the three gentlemen, owners of private land, who have graciously granted us permission to enter their property: Mr. Oscar Baker of Highgate Falls, Mr. Euclide Duhamel of Swanton, and Mr. Louis Gregoire of Mallett's Bay.

DESCRIPTION OF OUTCROPS AND ROAD LOG

7½' Topographic Quadrangle Maps: St. Albans (Vt.) and Milton (Vt.)

Start from Perkins Geology Hall, UVM.
Proceed to Interstate 89 north to Exit 17 (Champlain Islands).
Exit for Route 2 West.

Mileage

00.0 Enter Route 2 West.

00.7 Cedar Hill Gift Shoppe on left. Park cars at gift shop and walk to top of hill to the south. Fossiliferous outcrops are beneath power-line.

Stop 1 - The rock exposed here is a coarse gray-buff to tan weathering gray sandstone. It is referred to as the Monkton Quartzite. Abundant fossils may be found on weathered surfaces. At this locality, the commonest fossils are trilobite fragments, mostly disarticulated thoracic segments, but recognizable olenellid cephala, probably of *Olenellus*, and dorypygid cranidia and pydidia, probably of *Bonnia*, also occur. The fossils are preserved as molds in the sandstone matrix. All are disarticulated but not badly abraded; these circumstances suggest sedimentation and burial in gently moving water. The Monkton Quartzite probably represents a strand-line deposit.

The first systematic description of the fossils from this locality is that of Kindle and Tasch (1948). Further descriptions, with some taxonomic revisions, were published by Shaw (1962).

Return to Interstate 89.

01.4 Take Interstate 89 north to Exit 21 (Swanton), Approximately 25 miles.

00.0 Enter Route 78 West to Swanton.

01.0 Turn left (south) onto Route 7 in Village of Swanton.

02.0 Cross bridge over Missisquoi River.

03.9 Turn left just south of farmhouse on left of highway. Enter lane through gate. The Kelly quarry is just south of this lane approximately 0.1 miles east of the gate.

Stop 2 - The Kelly quarry, described by Schuchert (1937, p. 1035) and Shaw (1954, p. 1041), exposes two rock types in the lower Parker Slate. The lower of these is a dark-greenish-gray weathering gray-green micaceous slate with some interbeds of buff weathering gray laminated fine to medium grained sandstone. Dolomite nodules occur near the top. The upper rock type is a tan-buff weathering light gray dolomite, exposed on top of the knoll above and to the east of the quarry face.

Poorly preserved trilobite fragments may be found in the slate; fragments and external molds of Kootenia are fairly common in the overlying dolomite. Shaw (1954, p. 1041) gave a check list for both horizons.

Dactyloidites asteroides has been reported by Schuchert (1937, p. 1035) and Shaw (1954, p. 1041) from the slates at the entrance to the quarry. Shaw (1955, p. 784) reported a somewhat different form, D. edsoni, from the same locality. Dactyloidites has been described and figured by Ruedemann (1934, p. 28-30, plates 4-6) who thought it was probably algal. D. edsoni was also described by Resser and Howell (1938, p. 210) who thought it was algal. Walcott (1998) thought Dactyloidites was a scryphozoan medusa. Hantzschel (1962, p. W240) and Harrington and Moore (1956, p. F159) consider Dactyloidites to be an unrecognizable form.

Return to Route 7 north.

04.3 Turn right (east) on paved road.

05.6 Stop sign. Turn left (north).

08.3 Turn left (just before steel bridge over Missisquoi River at Highgate Falls) and drive through Swanton Municipal Power plant property to private property owned by Mr. Oscar Baker. Follow the lane to the right and park cars by river. Outcrops are to the east.

Stop 3 - In contrast to the first two stops, the last is dominated by carbonates. This section in the Highgate gorge has been described by Raymond (1924, p. 459), Schuchert (1933, p. 373-377; 1937, p. 1067-1069), and in greater detail by Shaw and Clark (1968).

Schuchert (1937, p. 1070) interpreted a thick breccia in the gorge as a thrust breccia and hence recognized two formations here, separated by this inferred thrust. The upper he referred to the Highgate Formation and the lower to the Gorge Formation. Shaw (in Shaw and Clark, 1968, p. 381) did not recognize a thrust in this part of the section,

interpreting the breccia in question as debris from a submarine land slide, and hence he assigned to the Gorge Formation the strata that Schuchert had referred to the Highgate Formation here.

Several fossiliferous horizons have been noted in the gorge section but the faunas of only some have been described. Clark and Shaw (1968a; 1968b) described the trilobites from bed 3, which is exposed downstream from the most westerly vertical cliff. This bed has yielded two distinct faunas, a lower referred by Clark and Shaw (1968a) and Palmer (1971, p. 176) to the late Dresbachian Dunderbergia zone, and an upper correlated by Clark and Shaw (1968b) with the Hungaia magnifica fauna, known from boulders in Quebec and western Newfoundland (Whittington, 1966, p. 701). Palmer (1971) referred the upper fauna in bed 3 to the late Franconian.

Higher strata have yielded fossils from a number of horizons. The lowest of these is stratigraphically about a foot above bed 3 and is exposed only in the same general locality as bed 3. Its trilobites have been described by Raymond (1924; 1937). Other fossiliferous horizons have been identified by Shaw and Clark (1968) in the most westerly vertical cliff section and in the cliff section to the east, between the two rock dumps. The fossils from some of these have been described by Raymond (1924; 1937) and Rasetti (1946), and have been correlated with the Hungaia magnifica fauna to be Trempealeauan and also correlative of the Early Tremadocian.

00.0 Return to steel bridge at Highgate Falls. Cross bridge.

00.4 Turn left in Highgate Center onto Route 78 west.

04.6 Enter Interstate 89.

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Whittington, H.B., 1966, Phylogeny and distribution of Ordovician trilobites: Jour. Paleontology, v. 40, p. 696-737.

APPENDIX

Vermont Geological Survey Publications

All Vermont Geological Survey Publications may be purchased through the Vermont Department of Libraries, Geological Publications, Montpelier, Vermont 05602.

Please include payment with your order. Vermont residents must include 3% sales tax.

*** VERMONT GEOLOGICAL SURVEY BULLETINS ***

1	Geology of the Bradford-Thetford Area, Orange County Vermont by Jarvis B. Hadley, 1950	2.00
2	Stratigraphy and structure of the Castleton Area, Vermont, by Philip Fowler, 1950	2.00
3	Geology of the Memphremagog Quadrangle and the Southeastern Portion of the Irasburg Quadrangle, Vermont, by Charles G. Doll, 1951	2.00
4	A Study of Lakes in Northeastern Vermont, by John Ross Mills, 1951	2.00
5	The Green Mountain Anticlinorium in the Vicinity of Rochester and East Middlebury, Vermont, by Philip Henry Osberg, 1952	2.00
6	The Geology of the Rutland Area, Vermont, by W.F. Brace, 1953	2.00
7	The Geology of the Bennington Area, Vermont, by John A. Mac-Fayden, 1956	2.00
8	The Geology of the Lyndonville Area, Vermont, by John G. Dennis, 1956	2.00
9	The Geology of the Limestone of Isle LaMotte and South Hero Island, Vermont, by Robert B. Erwin, 1957	2.00
10	The Bed Rock Geology of the East Barre Area, Vermont by Varansi Rama Murthy, 1957	2.00
11	The Geology of Concord, Waterford Area, Vermont by Eric and Dennis, 1958	2.00
12	The Geology of the Mount Mansfield Quadrangle, Vermont by Robert A. Christman, 1959	2.00
13	The Geology of the St. Johnsbury Quadrangle, Vermont and New Hampshire, by Leo M. Hall, 1959	2.00
14	Bedrock Geology of the Central Champlain Valley of Vermont, by Charles W. Welby, 1961	4.00
15	Geology of the Camels Hump Quadrangle, Vermont by Robert A. Christman and Donald T. Secor, Jr., 1961	2.00
16	Geology of the Plainfield Quadrangle, Vermont by Ronald H. Konig, 1961	2.00
17	The Green Mountain Anticlinorium in the Vicinity of Wilmington and Woodford, Vermont, by James William Skehan, S.J., 1961	3.00

*** VERMONT GEOLOGICAL SURVEY BULLETINS ***

18	Geology of the Equinox Quadrangle and Vicinity, by Philip C. Hewitt, 1961	2.00
19	The Glacial Geology of Vermont, by David P. Stewart, 1961	2.00
20	Geology of the Island Pond Area, Vermont, by Bruce K. Goodwin, 1963	2.00
21	Bedrock Geology of the Randolph Quadrangle, Vermont, by Ernest Henry Ern, 1963	2.00
22	Geology of the Lunenburg-Brunswick-Guildhall Area, Vermont, by Warren I. Johansson, 1963	2.00
23	Geology of the Enosburg Area, Vermont, by John G. Dennis, 1964	2.00
24	Geology of the Hardwick Area, Vermont, by Ronald H. Konig and John G. Dennis, 1964	2.00
25	Stratigraphy and Structure of a Portion of the Castleton Quadrangle, Vermont, by E-an Zen, 1964	2.00
26	Geology of the Milton Quadrangle, Vermont by Solon W. Stone and John G. Dennis, 1964	2.00
27	Geology of the Vermont Portion of the Averill Quadrangle, by Paul Benton Myers, Jr., 1964	2.00
28	Geology of the Burke Quadrangle, Vermont, by Bertram G. Woodland, 1965	3.00
29	Bedrock Geology of the Woodstock Quadrangle, Vermont by Ping Hsi Chang, Ernest H. Ern, Jr., and James B. Thompson, Jr., 1965	2.00
30	Bedrock Geology of the Pawlet Quadrangle, Vermont, by Robert C. Shumaker and James B. Thompson, Jr., 1967	2.00
31	The Surficial Geology and Pleistocene History of Vermont, by David P. Stewart and Paul MacClintock, 1969	4.00

*** ECONOMIC GEOLOGY ***

Economic Geology No. 1 - A Report on Magnetic Surveys of Uultramafic Bodies in the Dover, Windham and Ludlow areas, Vermont, by Vincent J. Murphy, 1966 — 3.00*

Economic Geology No. 2 - Report on a Resistivity Survey of the Monkton Kaolin Deposit and Drill Hole Exploration, by Jason A. Wark, 1968 — 3.00*

Economic Geology No. 3 - Geology and Origin of the Kaolin at East Monkton, Vermont, by Duncan G. Ogden, 1969 — 3.00*

Economic Geology No. 4 - Report on the Cuttingsville Pyrrhotite Deposit, Cuttingsville, Vermont, by Charles G. Doll, 1969 — 3.00*

Economic Geology No. 5 - The Geology of the Elizabeth Mine, Vermont, by Peter F. Howard, 1969 — 3.00*

Economic Geology No. 6 - Magnetic Surveys of Ultramafic Bodies in the Vicinity of Lowell, Vermont, by Vincent J. Murphy and Andrew V. Lacroix, 1969 — 3.00*

Economic Geology No. 7 - Geochemical Investigations in Essex and Caledonia Counties, Vermont, by Raymond W. Grant, 1970 — 3.00*

*** ENVIRONMENTAL GEOLOGY ***

Environmental Geology No. 1 - Geology for Environmental Planning in the Barre-Montpelier Region, Vermont, by David P. Stewart, 1971 — 2.00

*** SPECIAL PUBLICATIONS ***

Special Publication No. 1 - Paleontology of the Champlain Basin in Vermont, by Charles W. Welby, 1962 — 3.00

Special Publication No. 2 - Mineral Collecting in Vermont, by R. W. Grant, 1968 — 3.00

*** STUDIES IN VERMONT GEOLOGY ***

Studies in Vermont Geology No. 1 - The Morphometry and Recent Sedimentation of Joe's Pond, West Danville, Vermont, by John S. Moore and Allen S. Hunt, 1970 — 2.00

Studies in Vermont Geology No. 2 - Surficial Geology of the Brandon-Ticonderoga 15 Minute Quadrangles, Vermont, by G. Gordon Connally, 1970 — 2.00

*** MAPS ***

1	Topographic Map of Vermont, 1970, scale 1:250,000, contour interval 100'	2.00
2	Centennial Geologic Map of Vermont, 1961, scale 1:250,000	4.00
3	Surficial Geologic Map of Vermont, 1970, Scale 1:250,000	4.00
4	Generalized Geologic Map of Vermont, 1970, 8½ x 11" - each	.15
	In lots of 100 for schools - each	.10
5	Glacial Drift Sheets and Ice Directions - each	.15
	In lots of 100 for schools - each	.10
6	Post Card Generalized Geologic Map of Vermont, 1970, 4 7/16 x 6 7/16"	.10
7	Vermont Geological Quadrangle Maps - Areas Available: Castleton, Concord-Waterford, East Barre, Enosburg Falls, Equinox, Mt. Mansfield, Plainfield, Rutland, St. Johnsbury, Wilmington-Woodford - each	.25

*** STATE PARKS ***

Geology of Button Bay State Park, by Harry W. Dodge, Jr., 1962	.25
The Geology of Darling State Park, by Harry W. Dodge, Jr., 1967	.25
The Geology of Groton State Forest, by Robert A. Christman, 1956	.25
The Geology of Mt. Mansfield State Forest, by Robert A. Christman, 1956	.25
The Geology of the Calvin Coolidge State Forest Park, by Harry W. Dodge, 1959	.25
The Geology of D.A.R. State Park, Mt. Philo State Forest Park, Sand Bar State Park, by Harry W. Dodge, Jr., 1969	.25

*** SPECIAL BULLETIN

Special Bulletin No. 1 - Geology of the Plattsburgh and Rouses Point, New York-Vermont, Quadrangle by Donald W. Fisher, 1968	3.00

*** OTHER PUBLICATIONS ***

The Physical Features of Vermont, by Elbridge Churchill Jacobs, 1950	1.00

1830

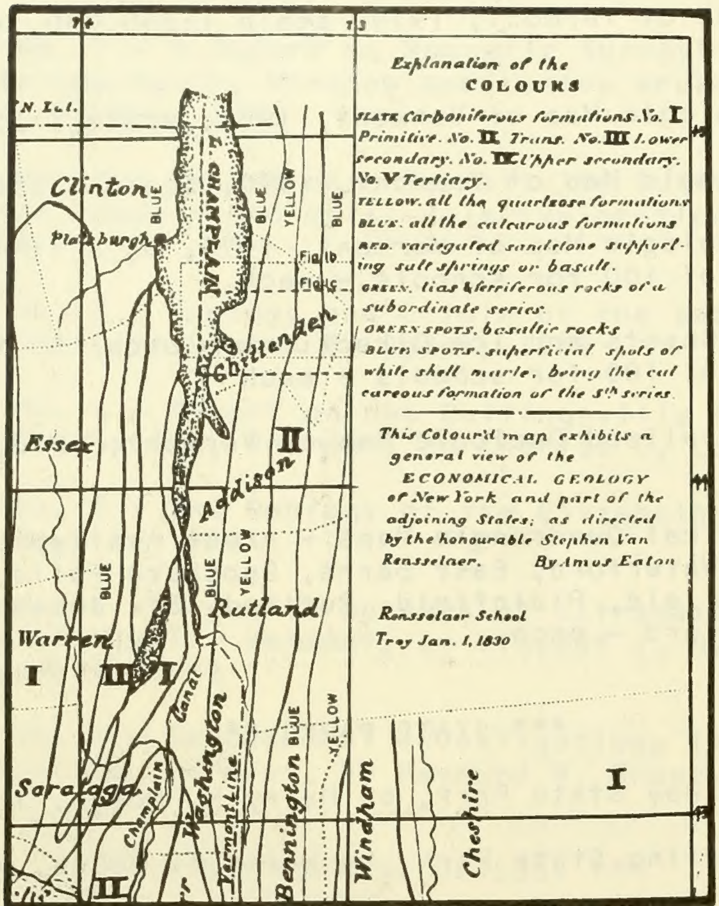

Fig. la

From A. Eaton (1830), Geologic Textbook, Pl. I

0 50 100 miles
Scale

1861

KEY

- (2) Gneiss
- (4) Talcose Schist
- (11) Chazy, Bird's Eye, and Black River Limestones
- (12) Trenton Limestone
- (13) Utica Slate
- (14) Hudson River Slates
- (16) Red Sandrock
- (17) Quartz Rock
- (18) Georgia Slates
- (19) Talcose Conglomerate
- (20) Eolian Limestone
- (S) Beds of Steatite
- (1) Ores of Iron older than the Tertiary

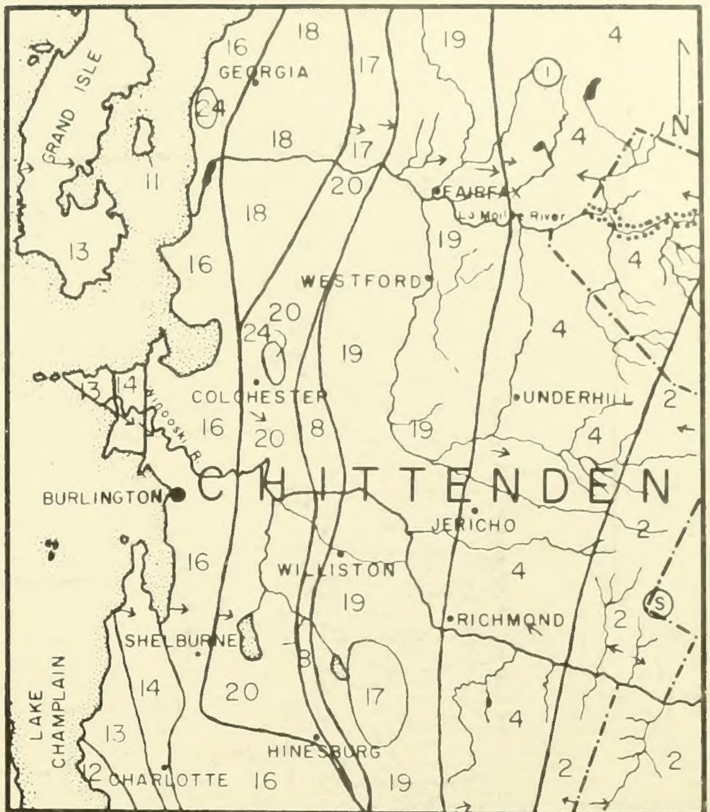

Fig Ib

Adapted from the E. Hitchcock's Report on the Geology of Vermont (1861), vol. 2, pl. 1

1961

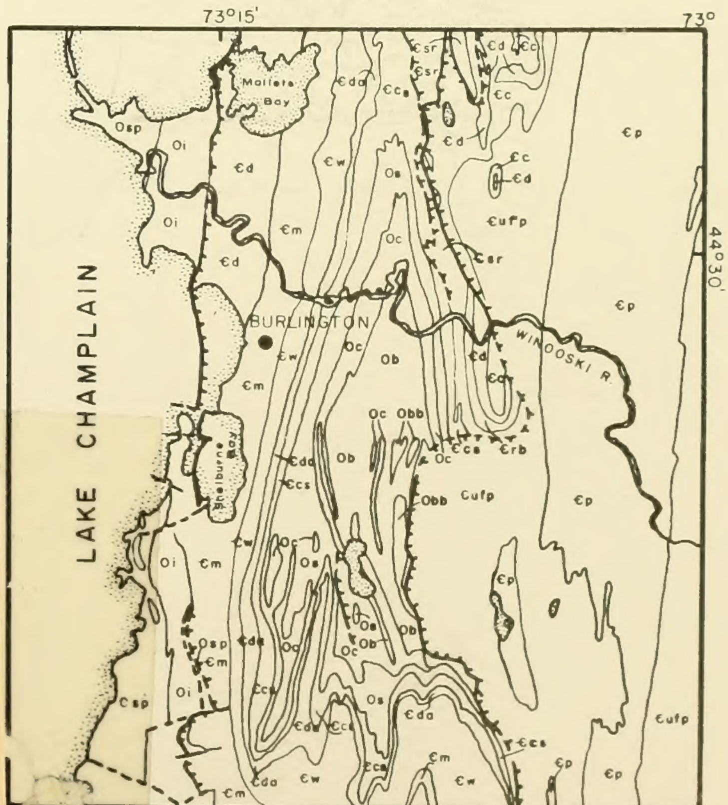

Fig. Ic

Adapted from the Centennial Geologic Map of Vermont by Doll et al. (1961).